

THE TECTONIC EVOLUTION OF THE
ROCKY CAPE GEANTICLINE
IN NORTHWEST TASMANIA

by

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This thesis contains no material which has been accepted for the award of any other degree or diploma in any university, and to the best of my knowledge and belief, contains no paraphrase of material previously published or written by another person except where due reference is made in the text of the thesis.



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A B S T R A C T

This thesis is a cross-sectional study of the Rocky Cape Geanticline, one of the major tectonic elements of Tasmania. The Rocky Cape Geanticline consists of deformed, unfossiliferous, and dominantly unmetamorphosed sedimentary rock of presumed Proterozoic age. The internal structure is due to the Penguin Orogeny of late Proterozoic age. Two distinct lithological assemblages are recognised, corresponding to different basins of deposition. These basins have a northeast-southwest trend, parallel to the axis of the Rocky Cape Geanticline, and flank the older Precambrian nucleus which lies to the east.

To the west of the axis of the geanticline is an unstable - shelf sequence about 20,000 feet thick, consisting of shale, siltstone, orthoquartzite and minor dolomite. This comprises the Rocky Cape Group; a revised definition is given for the Rocky Cape Group, and new names are given to the constituent formations. First record of sedimentation indicates a starved, euxinic environment. This was followed by a stable, shallow-water, free-circulating environment in which unusually large thicknesses of cross-bedded orthoquartzite accumulated. Palaeocurrent directions from cross-bedding have diametrically opposed, bimodal or polymodal patterns, which are interpreted as the expression of transverse dispersal currents. These may reflect a gently oscillating palaeoslope. Further to the west, the Rocky Cape Group is overlain, with a low-angle unconformity, by the Smithton Dolomite which has a basal conglomerate.

East of the axis of the geanticline, a flysch-type sequence at least 15,000 feet thick accumulated. This is called the Burnie Formation, and is considered to be younger than the Rocky Cape Group. It consists of slate, quartz wacke and minor pillow lava. Axially directed turbidity currents were an important agent of sedimentation,

although, large-scale festoon cross-bedding indicates that traction movement by bottom currents was significant. These currents are considered to be independent of the turbidity currents.

A variety of sedimentary deformation structures occur. Most are post-depositional, and can be further classified into those due either to down-slope mass movement, (block slide, slump sheet, and fluxoturbidite); or to small vertical, horizontal, or toroidal movements, (load cast, pseudonodule, diapiric contortion, ball-and-pillow structure and convolute lamination). Liquefaction is considered to be an important process in the development of the latter category. The imposed tectonic cleavage is commonly controlled by the pre-existing sedimentary deformation, but cross-cutting relationships also occur, and assist in the distinction between sedimentary and tectonic structures.

Major tectonic movements during the Penguin Orogeny were to the southeast, so that the Burnie Formation was squashed between the older basement to the east and the Rocky Cape Group to the west. Albite dolerite (Cooee Dolerite) in the form of dykes, sills and sheets were intruded during the early stages of folding. The two main sedimentary assemblages are now separated by the Keith Metamorphics, a belt about five miles wide, composed of pelitic schist, basic schist and amphibolite. The western boundary is gradational across strike, and displays cataclastic textures. The Keith Metamorphics was derived in part from the surrounding sediments and partly from basic igneous rock, believed to be equivalent to the Cooee Dolerite. This belt is interpreted as a high-angle shear zone. Correlation with similar schists along the axis of the Rocky Cape Geanticline indicates that this is a fundamental structure, and it is here termed the Arthur Lineament.

The structure in the Rocky Cape Group is simple, consisting of a series of folds trending parallel to the major axis. The folds are tight and asymmetrical near the shear zone, and become more open and symmetrical to the west. This folding affects the Smithton Dolomite.

Detailed structural analysis of the Burnie Formation reveals a complex picture of progressive deformation involving five phases, all on approximately the same axis. The first and third generations were the major events in the structural evolution. The major structure consists of two overturned synclines separated by a flat belt and an anticlinal hinge. The overall vergence is to the southeast, toward the older Precambrian nucleus.

Mesoscopic folds and associated structures of one phase tend to form in zones which are spatially segregated from zones of other generations. Thus, the first generation folds are mostly confined to the overturned limbs of the major synclines, and the third generation folds are confined to the intervening flat belt. Smaller scale segregation is demonstrated by the occurrence of the later phases in zones of planar bedding between clusters of first generation folds. The area of overlap of these zones is generally small, so that examples of refolded structures are not abundant, and a false impression is given of rapid changes in tectonic style of the one fold generation. The segregation of fold phases is most strongly expressed where the bedding remained kinematically active, a feature which suggests that there is a mechanical limitation to superposed folding.

In the first generation mesoscopic folds, the geometry of the arenite beds approximately fits that of flattened concentric folds. At Sulphur Creek, the mesoscopic folds have a marked asymmetry, shown by common-limb thinning of the fold couplet, a "half-fan"

cleavage, and offset carinate structures. These features may be explained by an obliquity between the initial axial plane and the plane of flattening. The flattening stage of fold development was followed by a disharmonic stage involving variable lateral shortening, and the formation of "slip-off" structures, oblique shear joints and boudinage. Cleavage formed continuously throughout the sequence of development of the P1 minor structures, although the main period of formation was during the flattening stage. The change in behaviour of the arenite beds during folding from initially competent, to incompetent, and back to competent, may be due to the role of internal pore water during deformation.

The evolution of the Rocky Cape Geanticline fits into a pattern of Proterozoic - Lower Palaeozoic orogenesis, in which the axis of sedimentation progressively shifted toward the older nucleus, and the tempo of sedimentation and structural deformation increased to a climax in the Upper Cambrian.

CHAPTER 1

INTRODUCTIONSCOPE

This thesis outlines the sedimentary and structural history of the Rocky Cape Geanticline in northwest Tasmania. In effect, it is a cross-sectional study of the axial portion of a deformed Precambrian sedimentary basin.

The Rocky Cape Geanticline, one of the major tectonic elements of Tasmania, is an elongate basement high, extending across the northwest corner of Tasmania. It is composed mainly of unmetamorphosed sedimentary rocks. The axis extends from about Trial Harbour on the west coast to Wynyard on the northwest coast, a distance of 70 miles.

Another geanticlinal high, called the Tyennan Nucleus, flanks the Rocky Cape Geanticline about 40 miles to the east. This element forms the nucleus of Tasmania. It is composed of metasediments which, by definition (Spry, 1962) record effects of the Frenchman Orogeny. These rocks are generally considered to be older than those unmetamorphosed rocks which make up the Rocky Cape Geanticline. The deformation recorded in these unmetamorphosed rocks of the Rocky Cape Geanticline is attributed to the Penguin Orogeny (Spry 1962, p.124), of late Precambrian age.

The geosynclinal trough between these two positive elements was the site of lower Palaeozoic sedimentation, orogenesis and important metallogenesis. Consequently, this trough has received considerable geological attention. However, the Rocky Cape Geanticline has received little attention and its stratigraphical and structural framework is virtually unknown.

The aim of this thesis is primarily to outline the evolution of the Rocky Cape Geanticline in northwest Tasmania through the sedimentational and deformational stages. It is intended, by the demonstration of a major coherent structure, to fill a substantial gap in the knowledge of the tectonic history of Tasmania.

This study also has an important bearing on some basic concepts on subdivision of the Precambrian rocks of Tasmania. The age relationships between the metamorphosed and unmetamorphosed rocks is one of the major problems of Tasmanian Precambrian geology. Generally, the metamorphosed rocks are considered to be older than the unmetamorphosed rocks. This area provides an example where the unmetamorphosed rocks are in gradational contact with a strip of metamorphosed rocks lying along the axis of the geanticline. It is suggested that although the above approach to age sub-division may be generally correct, it does not

apply in the present case. The distinction should therefore be made on structural as well as petrological grounds.

The constancy of trends and the presence of a major structure, allow some extrapolation of the basic relationships along the axis of the geanticline to the southwest into the west coast mineral belt. It appears that some of the previous ideas on the configuration of the upper Precambrian in western Tasmania are in need of revision.

This thesis has a strong bias toward structural geology because of the presence of a repeatedly deformed flysch-type sequence in the vicinity of Burnie. Some considerations are made on the techniques of mesoscopic structural analysis, and considerable attention is given to the geometry and mechanics of folding and cross-folding.

Along the northwest coast of Tasmania, the structural grain trends northeast-southwest perpendicular to the coast line, providing a well exposed cross section of the Rocky Cape Geanticline.

The area studied covers a large portion of the northwest coast of Tasmania, from Penguin westward almost to Smithton, and extending inland for up to 16 miles. This work was carried out in conjunction with the regional mapping of the Burnie and Table Cape

Quadrangles by the Geological Survey of the Tasmanian Department of Mines.

Table Cape Quadrangle has been published and is included here as Figure 59, and Burnie Quadrangle is in press. However, for the purpose of this study no specific area was delineated, and some of the investigation also falls into parts of Devonport, Smithton and Trowutta Quadrangles.

Figure 1 is a locality map showing the major towns, and the extent of the area investigated. Between Penguin and Wynyard excellent exposure is found on a well-developed shore platform. The mean rise and fall of tide is 12 feet. The Bass Highway and the Western Railway follow the coastline along the entire section. Inland, there are numerous country roads which serve the rich farming district, but the exposure is limited by extensive sheets of Tertiary basalt. Between Wynyard and the Black River both access and exposure is reasonably good. Here the Bass Highway and the Western Railway turn inland to cross the Sisters Hills. In the south and west of the area there are some dry-weather timber roads which provide sufficient access into the Sisters Hills, Shakespeare Hills, Dip Range and Campbell Range.

The Rocky Cape Geanticline is composed of two thick sedimentary sequences. In the eastern half

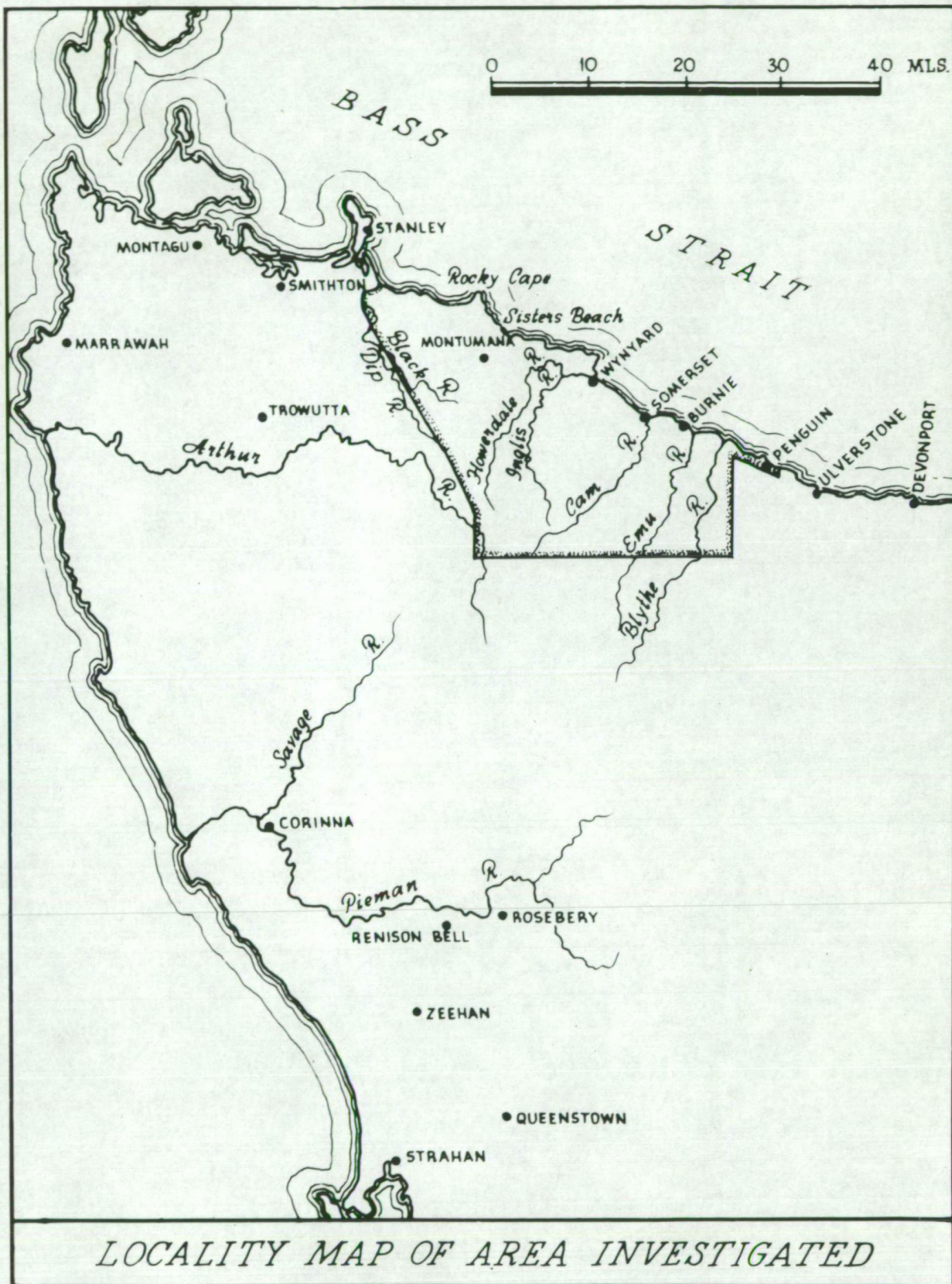


Figure 1

of the area between Penguin and Wynyard, is a repeatedly deformed, flysch-type sequence. The shore platform has been mapped on a scale of 200 feet to one inch on small-scale aerial photographs kindly supplied by Dr. K. L. Burns and by the Department of Mines. The detailed maps are presented as Figures 54, 55, 56, 57 and 58 in the folder at the back.

The major fold hinges have been extended inland for up to seven miles by structural mapping in the river gorges.

Further to the west, between Wynyard and Black River there is an unstable shelf sequence consisting mainly of orthoquartzite, siltstone and shale. Approximately 200 square miles in this sequence has been mapped in order to demonstrate a major coherent structure and a stratigraphic succession. No base has been found in this succession.

Separating these two deformed sedimentary sequences is a thin belt of metamorphic rocks which is interpreted as a high-angle shear zone. This belt is considered to be the direct continuation of the Whyte Schist (Spry, 1964) at the southwestern end of the geanticline. It defines a lineament which is the most important structure in northwest Tasmania. Spry (1964, p.44) envisaged the belt of metamorphic rocks as an older basement high which was metamorphosed in the Frenchman Orogeny, with younger sediments deposited

in flanking basins. The metamorphic belt is here considered to be a fundamental shear zone of Proterozoic age that formed during the Penguin Orogeny.

In this thesis, specimen numbers refer to catalogue numbers in ^{the} collection of the Geology Department, University of Tasmania. All bearings given are relative to true ~~grid~~ north. All orientation data are given on equal-area projections, except where otherwise stated.

HISTORICAL REVIEW

The first geological recording of the rocks of the northwest coast of Tasmania was by Strzelecki (1845, p.93) who observed contorted siliceous and argillaceous slates west of Emu Bay at Burnie. Much later, Stephens (1870) noted these folded rocks, and also the intrusive dolerite bodies. These dykes attracted prospectors and traces of copper were found in many places between Rocky Cape and Penguin. Hematite was also discovered in the Blythe River.

Montgomery (1894) visited the area to examine the iron ore at Blythe River, and again in (1896) to examine some of the alluvial gold fields near Wynyard. Montgomery also mentioned the quartzites at Rocky Cape, and appears to be the first to note the highly micaceous crystalline schists in the Inglis River.

Twelvetrees (1903, 1905) made the first worthwhile observations in reports on the mineral fields of the northwest coast. He broadly referred to the rocks west of Wynyard as the "Rocky Cape quartzites", which he considered to be Precambrian. He also described the strongly contorted slate and quartzite near Burnie which he considered was probably Cambro-Ordovician. Once again, Stephens (1909) in a journey from Tamar River to Circular Head made the first mention of conglomerate and limestone (dolomite) at Black River. Hills (1913) also visited the area to assess the coal potential of the Permian rocks in the Wynyard area.

With the failure of the mineral fields, the northwest coast of Tasmania received little further geological attention. Nye, Finucane and Blake (1934) carried out systematic mapping in the Smithton area. These authors noted the quartzite and the widespread occurrence of dolomite.

Following the delineation of the west coast mineral belt, the barren rocks to the northwest began to emerge as an important structural element, although their stratigraphic position remained unknown. Hills and Carey (1949) variously assigned them to the Cambrian and to the Precambrian. This uncertainty persisted to 1953 when Carey recognised and defined the Rocky Cape

Geanticline (1953, p.1115). With the recognition of Middle Cambrian sediments and volcanics in the west coast mineral belt (Scott, 1954, Bradley, 1954 and Elliston, 1954), the rocks of the Rocky Cape Geanticline were then thought of as upper Precambrian and possibly extending into the lower part of the Cambrian.

In the light of this new work, Carey and Scott (1952) had already given a revised interpretation of the Smithton district. These authors considered that the basement rocks were late Precambrian and recognised that the quartzite at Bryant Hill was overlain by the Smithton Dolomite which in turn was overlain by a meridional belt of Cambrian spilite. This was confirmed by Gulline (1959), who found Middle Cambrian fossils in an assemblage of siltstones and volcanics lying above the Smithton Dolomite. This permitted a precise time correlation with the Dundas Group at Zeehan. Gulline also noted the conglomerate at Black River.

The work of Spry (1957a) was the first attempt at stratigraphic subdivision within the Rocky Cape Geanticline in northwest Tasmania. He defined the Rocky Cape Group to include all presumed upper Precambrian rocks between Penguin and Smithton and set up some formational names. He suggested a succession consisting of Black River Dolomite at the base, followed by Cowrie Siltstone, Bluff Quartzite, Port Slate, and

Cave Quartzite. The Burnie Quartzite and Slate was thought to be older than all these.

Spry (1962, 1964) modified his original stratigraphic succession when it was recognised that the conglomerate at Black River overlies the Cowrie Siltstone. This thin conglomerate horizon was then interpreted as a facies variant of the Bluff Quartzite which also overlies the Cowrie Siltstone, eight miles to the east at Rocky Cape. This correlation is not supported by the present work. Spry continued to envisage the Burnie Quartzite and Slate beneath all the others, and again this is not supported. In addition, some of the formations are found not to be mappable units, whereas there are other mappable units above the Cave Quartzite of Spry. The problems of stratigraphic terminology are discussed in Chapter 2 and the Rocky Cape Group is redefined.

Reconnaissance mapping was carried out in an adjacent area in the middle portion of the Arthur River by McNeil (1964), who suggested a sequence of Keith Beds (metamorphic rocks) at the base, followed by Neasey Quartzite and Lawson River Siltstone. Present work indicates that the sequence is inverted. The Neasey Quartzite encompasses the three top formations of the revised Rocky Cape Group. Longman and Matthews (1963) also did reconnaissance mapping further to the west. These two recent contributions showed that the

fold trends in the Rocky Cape Geanticline were directed northeast-southwest.

Following his work on the lower Pieman River area, Spry (1964) suggested some regional correlations along the length of the Rocky Cape Geanticline. The more important of these were the correlation of the Whyte Schist with the Keith Beds, the correlation of the Oomah Quartzite with the Burnie Quartzite and Slate. These steps are followed. Other suggested correlations which are not substantiated are the equivalence of the Burnie and the Neasey, and the equivalence of the Keith Beds and Ulverstone Metamorphics of Burns (1964).

The eastern border of the Rocky Cape Geanticline has recently received detailed attention from Burns (1964), who described a complex trough of Cambrian sediments and volcanics between the geanticline and the metamorphic basement to the east at Forth. Burns believed that the unmetamorphosed Burnie Formation was thrust onto the metamorphic Forth Nucleus. This concept is supported but the suggestion that the thrust may be the expression of telescoping of metamorphic facies within Precambrian rocks of equivalent age is not followed.

The precise age of the sediments of the Rocky Cape Geanticline remains uncertain. The Precambrian of Tasmania has been divided into metamorphic and non-

metamorphic rocks, and Spry (1962, p.120-126) has outlined the evidence that this also represents an older and younger division. Although the argument is inconclusive, this approach is followed as a working hypothesis. One additional point of supporting evidence is that the rocks of the Rocky Cape Geanticline constitute almost a discrete sedimentological and structural entity within itself.

The Burnie Formation, considered here to be the youngest of the rocks in the Rocky Cape Geanticline, is older than upper Middle Cambrian siltstones at Penguin, (Burns, 1964). It is thus possible that some of the rocks may extend up into the Lower Cambrian.

All the rocks have so far proved to be unfossiliferous. It is possible that stromatolithic-type growths will be found in the Smithton Dolomite as in other Proterozoic dolomites elsewhere in the world. Crinoid ossicles have been reported from the Smithton Dolomite by Chapman (in Nye, Finucane and Blake, 1934, p.39), however these can neither be substantiated or confirmed. A single radiometric dating on the Cooe Dolerite which intrudes the Burnie Formation at Burnie indicates an age older than 700 million years for the Burnie Formation, Spry (1962). With these reservations in mind the sedimentary rocks of the Rocky Cape Geanticline are referred to in this thesis as Proterozoic.

ACKNOWLEDGEMENTS

This work has been carried out while on the staff of the Geological Survey of Tasmania, and the writer is grateful to the Director of Mines for granting of two years leave of absence to attend the University of Tasmania. All possible assistance and encouragement has been given by the Mines Department who supplied assistance in the field, gave assistance in drafting, carried out six chemical analyses and furnished small-scale aerial photos. In particular the writer would like to convey his appreciation to Dr. E. Williams who has shown a keen interest and tendered much helpful advice throughout all the stages of the work. The author is indebted to Professor S. W. Carey of the University of Tasmania and to Dr. A. Spry who supervised, giving valuable guidance and criticism. The writer has benefited from frequent discussions from fellow graduate students, particularly Mr. C. Powell.

CHAPTER 2

STRATIGRAPHICAL TERMINOLOGYINTRODUCTION

Stratigraphical terminology in the Proterozoic succession in northwest Tasmania is mainly after Spry (1957a) who introduced several formation names. The sequence proposed was later modified, (Spry, 1962, p.111) as follows :

Top - Cave Quartzite
Port Slate and Quartzite
Bluff Quartzite
Cowrie Siltstone
Base - Burnie Quartzite and Slate

These formations were taken to constitute the Rocky Cape Group which was defined (Spry, 1957a, p.81) as "those sediments chiefly quartzite slate dolomite and siltstone, outcropping intermittently from Penguin to Smithton, and lying unconformably below the Dundas Group at Penguin. Its thickness is in excess of 10,000 feet."

In addition, other formations have been named in northwest Tasmania, for example, Bryant Hill Quartzite (Carey and Scott 1952), Smithton Dolomite (Spry 1957b), Black River Dolomite (Spry 1957a), and Forest Conglomerate and Quartzite (Spry 1964, p.47). Although these formations fall within the definition of the Rocky Cape Group, it is not clear, either from the original definitions or from current usage, whether

they should be included in the Rocky Cape Group.

It can now be shown that the succession between Penguin and Smithton consists of three assemblages of contrasting lithological characteristics, each assemblage corresponding to a different basin of deposition. It is therefore unjustified to call this succession a group in accordance with the Australian Code of Stratigraphic Nomenclature (1964). Furthermore, there is a proved angular unconformity separating two of these assemblages.

Figure 2 summarises the stratigraphical relations of the rocks comprising the Rocky Cape Geanticline.

There are three alternatives available to eliminate the difficulties of present nomenclature.

(1) The Rocky Cape Group, taken to include all rocks of supposed upper Proterozoic age between Penguin and Smithton, may be divided into sub-groups to correspond with the main lithological entities. This procedure does not rectify the incorrect use of the word "group".

(2) The Rocky Cape Group may be raised in status to a super-group, and new groups defined, a procedure suggested in the Australian Code of Stratigraphic Nomenclature (1964). This is unwise since the use of the term "Rocky Cape" will become broader as knowledge

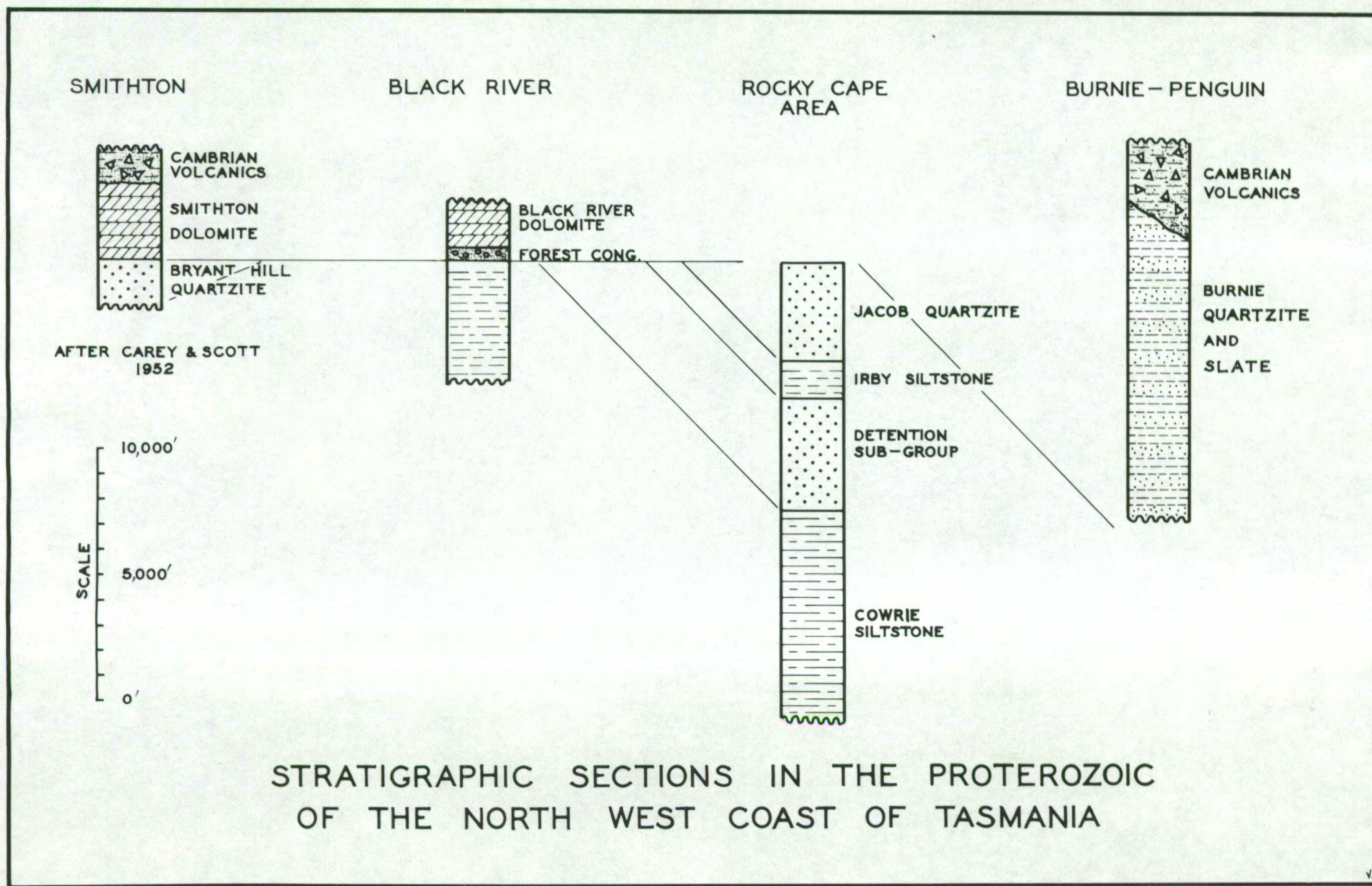


Figure 2.

of the regional geology increases. The writer is in favour of "Proterozoic succession" as the all-embracing term for those presumed upper Proterozoic sediments in northwestern and western Tasmania.

(3) The Rocky Cape Group may be restricted in definition to include only those quartzites and siltstones, some of which have been defined by Spry (1957a), which form a continuous sequence in the Rocky Cape and Dip Range area. This would include all those formations listed previously on page 15, with the exception of the Burnie Quartzite and Slate. It would also exclude the Smithton Dolomite, the Black River Dolomite and the Forest Conglomerate which unconformably overlies the sequence in question. This procedure has several advantages; it retains much of the meaning of the original definition, it is a valid use of the word "group" permitting subdivision into sub-groups, and it does not interfere with the naming of hitherto undiscovered groups within the Proterozoic succession. The Rocky Cape Group is redefined below.

ROCKY CAPE GROUP

The Rocky Cape Group is that group of rocks, mostly quartzite, siltstone and mudstone, at least 16,000 feet thick, consisting of the Cowrie Siltstone (at the bottom), Detention Sub-Group, Irby Siltstone,

Jacob Quartzite (at the top), and including any other formation which can be shown to be conformably above the Jacob Quartzite or conformably below the Cowrie Siltstone. It occurs in the general area around Rocky Cape, Sisters Hills, Dip Range, Mawbanna; and is unconformably overlain to the west by the transgressive Smithton Dolomite.

The Rocky Cape Group consists of one sub-group and three additional formations.

Cowrie Siltstone

Spry (1957a) proposed the name Cowrie Siltstone for these well-bedded siltstones which outcrops along the foreshore between Rocky Cape and Black River. At that time its stratigraphic position was not understood, but later Spry (1962, 1964) recognised that it lay beneath the orthoquartzite at Rocky Cape. In accordance with the usage of Spry, it is formally defined as follows:

The Cowrie Siltstone is that formation of finely laminated shale, well-bedded flaggy or laminated siltstone, and black mudstone, with some cross-bedded sandstone layers, lying conformably beneath the Detention Sub-group; it is approximately 8,000 feet thick, and its type locality is on the western foreshore of Rocky Cape (966,300 y N, 332,500 y E), 1½ miles from the point of Rocky Cape. It occurs extensively on the flat plains between Sisters Hills and Black River, and also in the head-waters region of the Black and Dip Rivers.

Its outcrop is limited because of the cover of Tertiary basalt and Recent sediments. It outcrops extensively along the foreshore in the vicinity of Cowrie Point.

The thickness of the Cowrie Siltstone is difficult to estimate because of a large transcurrent fault, and the effects of folding. The figure of 8,000 feet is an estimate based on an assumed regional dip with a correction for the transcurrent movement. The fold profiles at Cowrie Point indicate that the folding had little effect on the thickness.

Detention Sub-group

The Detention Sub-group is proposed to embrace the Bluff Quartzite, Port Slate, and Cave Quartzite of Spry (1957a). Mapping has shown that these three units taken together constitute the only mappable unit. The Port Slate of Spry is only 200 feet thick and cannot be traced for more than a few hundred yards. Thin, non-persistent lenses of siltstone comprise about 15% of the Detention Sub-group.

The Detention Sub-group is defined as that assemblage of dominantly cross-bedded orthoquartzite with minor inter-bedded non-persistent siltstone beds, lying conformably above the Cowrie Siltstone at Rocky Cape and below the Irby Siltstone at Sisters Beach. It is 4,600 feet thick and has its type locality at Rocky Cape where it consists of the Bluff Quartzite, the

Port Slate and the Cave Quartzite.

The Detention Sub-group forms the dominant line of hills between Rocky Cape and Sisters Beach. One mile south of Sisters Beach, it is displaced five miles dextrally by a large transcurrent fault, to a point just north of Mawbanna. From here it can be followed around the Newhaven syncline, over the Sisters anticline to the Dip Range and then almost to the Arthur River, (Figure 16).

Irby Siltstone

This is defined as that formation of black siltstone and minor dolomite and sandstone, lying conformably above the Detention Sub-group and below the Jacob Quartzite; its type locality is at Sisters Beach. It is named after Irby Flats behind Sisters Beach.

It also occurs in the core of the Newhaven syncline near Montumana and also in the Dip Range where it can be followed almost to the Arthur River. The thickness is not precisely known because it is an incompetent formation sandwiched between two massive quartzite formations. At Sisters Beach it is about 2,500 feet thick.

Jacob Quartzite

This is defined as that formation of dominantly cross-bedded orthoquartzite with minor interbedded non-persistent siltstone horizons, being about 3,700 feet thick, and having its type locality on the headland at Jacobs Boat Harbour, after which it is named.

It outcrops continuously along the coast between Sisters Beach and Jacobs Boat Harbour, and also forms the syncline on the hills just inland from Jacobs Boat Harbour. In the core of this syncline, forming the contorted southern block of the major east-west transcurrent fault (Figure 16), is about 200 feet of siltstone and mudstone overlying the Jacob Quartzite. This un-named siltstone is taken as the top of the Jacob Quartzite.

SMITHTON DOLOMITE

The Rocky Cape Group is overlain in the west by a blanket of dolomite which in places has a basal conglomerate layer. The widespread occurrence of dolomite in the Smithton district was first noted by Nye, Finucane and Blake (1934), and has been referred to frequently by many writers, including Carey and Scott (1952), and Hosking and Heuber (1954). Spry (1957b) formally defined the Smithton Dolomite as, "The formation chiefly dolomite lying below the Dundas Group and above the Bryant Hill Quartzite of Carey and Scott (1952),

being approximately 3,000 feet thick, with its type locality being immediately west of the Duck River, just north of the Smithton-Marrawah Road."

Both Longman and Matthews (1962), and Spry (1962, p.112, 1964 p.47) noted the regionally transgressive nature of the Smithton Dolomite and have suggested an unconformity at the base.

The dolomite at Black River was named by Spry (1957a) the Black River Dolomite, who thought it was low down in the Rocky Cape Group. Later, Spry (1964) ascertained that the Cowrie Siltstone at Black River was overlain by a conglomerate, quartz sandstone, chert, and then the Black River Dolomite. The Black River Dolomite is correlated with the Smithton Dolomite by lithological identity and close proximity to the Smithton Dolomite near South Forest and Irishtown, (Gulline 1959). Spry (1964) named the conglomerate and quartz sandstone that underlie the dolomite at Black River the Forest Conglomerate and Quartzite after the nearby village of Forest.

Figure 17 is a geological map of the Black River estuary, and the stratigraphic position of the rocks is shown in Figure 2.

The sequence at Black River is as follows :

Black River (Smithton) Dolomite	30 feet
banded chert	30
quartz sandstone with cross-bedding	20
conglomerate with minor sandstone	25
unconformity	
Cowrie Siltstone	8,000 +

In the Black River area there is an angular discordance between the Cowrie Siltstone and the Forest Conglomerate. This discordance is as much as 70° of strike and 30° of dip. The unconformity is best exposed in the bank of the Black River, 1 mile upstream from the new road bridge, at the locality shown on Figure 17. Here the conglomerate dips $065^{\circ}/60^{\circ}$ NW and the siltstone dips $090^{\circ}/61^{\circ}$ N, giving an angular discordance of 22° . The base of the conglomerate transgresses across four feet of siltstone in a width of exposure of 15 feet.

The best indication of an unconformity is the presence of rectilinear grooves and ridges on the sole of the basal conglomerate bed, Plate 1-a.

Superficially these ridges resemble incipient rotational or shear joint boudinage, however there are no joints in the conglomerate or siltstone having the

required orientation. The ridges are the casts of sand and fine-pebble infillings of depressions between small strike ridges on an erosional surface which truncates the bedding at an acute angle. The sharpness of these ridges indicates that the siltstone must have been well compacted prior to the erosion. It is noteworthy that the cast ridges are present only when the size of the pebbles is less than the spacing of the strike ridges.

KEITH METAMORPHICS

The Rocky Cape Group is bounded on the east by a belt of low-grade greenschists. This belt can be followed from Wynyard southwest through Meunna to the Arthur River, and is a direct continuation of the Keith Beds of McNeil (1961). This ribbon-like belt is about four miles wide, and separates the Rocky Cape Group and the Burnie Formation. The schists are considered to be derived mainly from the surrounding sediments and partly from intrusive dolerite. The western contact, although gradational, is parallel to bedding in the adjacent Rocky Cape Group. These formations strike northeast with steep dips facing the southeast and in places are overturned. Thus the metamorphic rocks stratigraphically overlie the Rocky Cape Group. It is defined below as a rock-unit, considered to have been derived from sedimentary rocks

which lie stratigraphically above the Jacob Quartzite.

The Keith Metamorphics is that assemblage of phyllite, muscovite schist, calcite schist, albite schist, albite-amphibole schist, amphibolite and quartzite, originally named by McNeil the Keith Beds; which forms a linear belt between Wynyard and the Arthur River near the Keith River, and probably extends much further to the southwest. It lies stratigraphically above the Rocky Cape Group, and its type locality may be taken as the area of best exposure which is where Hilder's timber road from Meunna meets the Arthur River.

BURNIE QUARTZITE AND SLATE

The Burnie Quartzite and Slate was defined (Spry 1957a, p.81) as the formation of sub-graywacke quartzite and slate outcropping along the foreshore at West Burnie, and which appears to outcrop from Howth to Doctors Rocks, except where covered by later superficial material.

Spry (1962, 1964) suggested that the formation was the lowest in the Rocky Cape Group and the general dip was toward the west. Detailed mapping between Penguin and Doctors Rocks has shown that large tracts are overturned, and that the formation is generally younging to the east. Therefore, the Burnie Formation appears to lie above the Rocky Cape Group, and to be separated from it by the Keith Metamorphics.

The Burnie Formation has a minimum thickness of 14,500 feet. It is unlike the Rocky Cape Group and was laid down in a different basin of deposition. Therefore it is not included in the Rocky Cape Group. It may justifiably be called a group, but it is retained with formational status because most investigation has been centred on the shore platform. Generally, it is lithologically homogeneous, but there are some horizons up to 600 feet thick with very little arenite, and at Penguin there is a more siliceous sandstone. These could be designated formations, but they could not be exploited as mappable units. Future mapping may show that the Burnie Formation does contain mappable units and then it should be raised in status to a group.

PALAEOZOIC ROCKS

Palaeozoic rocks occur on the eastern and western flanks of the Rocky Cape Geanticline at Penguin and Smithton respectively. Although they do not contribute to the Rocky Cape Geanticline, the Palaeozoic tectonic movements have affected the Rocky Cape Geanticline. These sediments are therefore discussed only in regard to their structural history and not their sedimentary history. The following paragraphs summarise their stratigraphy.

Cambrian

The Burnie Formation at Penguin is overlain

by the Beecraft Megabreccia (Burns 1964, p.53) consisting of autochthonous graywacke and large allochthonousth blocks of varied but local origin. Burns believed it to be high in the Dundas Group in the Leipyge horizon of the Middle Cambrian. Burns considered that the Cambrian unconformably overlies the Burnie Formation on a high-angle transgressive contact, which was probably a fault scarp marking the western edge of the Dial Trough at Penguin.

The Cambrian stratigraphy is incompletely known at Smithton, but Carey and Scott (1952) and Gulline (1959) considered that the Smithton Dolomite is in part overlain by spilitic volcanics and pyroclastics, and then by siltstone that contains upper Middle Cambrian trilobites. The Cambrian is now confined to an ill-defined trough called the Montagu Synclinorium, (Banks 1959).

Ordovician

The Burnie Formation is overlain with marked angular unconformity by gently folded Ordovician Conglomerate. At Johnsons Beach, Penguin, it has been termed the Duncan Conglomerate. These are post-orogenic (Jukesian Orogeny) conglomerates of limited lateral extent, derived mainly from close-by Cambrian cherts which were originally deposited in the Dial Trough.

C H A P T E R 3

SEDIMENTARY FEATURES OF THE ROCKY CAPE GROUPCOWRIE SILTSTONE

The Cowrie Siltstone is exposed almost continuously along the coastline from Rocky Cape westward to the mouth of the Black River. This is more or less a strike section along which there is a general decrease in grain size to the west. At Rocky Cape, the rock-type is a coarse quartz siltstone with interbedded black shale. In the vicinity of Cowrie Point it is dominantly a massive or laminated black pyritic siltstone and shale, and further to the west at Black River it is a finely laminated shale. This regional variation is also apparent in the inland exposures.

Rocky Cape

On the west side of Rocky Cape, the dominant rock type is a coarse-grained, grey siltstone with bedding units from 6 inches to 30 inches thick. These siltstones (specimens 33276 and 33277) contain about 60% angular quartz grains 0.01 - 0.15 mm in diameter, about 5% twisted and ragged detrital muscovite, 2% of large (0.02mm) rounded flakes of detrital chlorite, and the remainder a chlorite-sericite matrix. Accessories are zircon, rutile, tourmaline and hematite. Also present are claystone fragments up to one cm long.

These coarse-grained quartz siltstones are interbedded with thickly bedded dark gray fine-grained siltstones, some of which have a fine internal lamination. These consist (for example, 33273) of quartz, muscovite and abundant large rounded flakes of chlorite (Plate 1-b) in a matrix of semi-opaque chlorite and sericite. Matrix amounts to about 50%. The laminations are due to finer claystone or coarser siltstone layers varying from a fraction of a mm up to one cm thick. These coarse-grained siltstone laminae display intricate distortions due to load deformation, and these are described in Chapter 5.

Cowrie Point

At Crayfish Creek and Cowrie Point the dominant rock type is a medium-grained siliceous siltstone, interbedded with black pyritic mudstone. The siltstone is regularly bedded in units ranging from 2 inches to 8 feet thick, and frequently possesses a faint internal lamination. Specimens 6155, 33281 and 33283 are typical siltstones, containing 40% medium to coarse silt-size quartz, 15% feldspar (albite and rare microcline), 3% clastic muscovite and the remainder a matrix of microcrystalline quartz, chlorite and sericite.

An unusual feature of some of the siltstone is the extremely high silica content, giving a glassy and cherty appearance in hand specimen. Spry (1964, p.46)

records an analysis which gives a silica content of 73.9% which is high compared with 58% for the average shale. (Pettijohn, 1957 p.106). In thin section, (6157, 33281) the matrix is seen as a mosaic of clear interlocking microcrystalline quartz, varying in size from 0.002mm to 0.015mm. These fine cherty grains have distinctively fuzzy outlines. Sericite is present in the matrix, but accounts for only 10% of the slide. Spry (1964, p.47) has previously noted these unusual textures. It is unlikely that the matrix silica is recrystallized detrital material because of the fine texture and the high silica content. There is no evidence of a biochemical origin. It is probable that it is a syngenetic chemical precipitate that has recrystallized during diagenesis. The origin of the silica is unknown. There are no known volcanic rocks in the basin that might supply syngenetic silica and the siliceous siltstone bears no relationship to the intrusive dolerite that occurs.

Interbedded with the siltstone is black, pyritic and well-bedded shale with bedding units averaging 2 feet in thickness. In thin section these consist of 30% quartz (0.01 - 0.002mm), fine crystalline sericite and abundant graphitic material. There is a minor amount of small (0.04mm) ragged detrital muscovite, and accessories include rutile, rounded tourmaline and detrital chlorite.

The black colour is due to the abundance of graphitic material and disseminated pyrite.

The pyrite occurs in cubic crystals with an edge-length of up to 10mm. It generally forms in films along bedding planes, or as nodules. The nodules are spherical, discoid or oblate in shape, varying in size from $\frac{1}{2}$ inch up to 6 inches in diameter. A few rare examples reach 5 feet in diameter. Where discoid, the greatest diameter is in the bedding plane, a feature which suggests preferential growth along certain laminae. Internally the nodules are composed of sedimentary rock rich in chlorite with a porous, brown-stained boxwork. The outer rim is rich in pyrite cubes, and also contains small quartz crystals and fibrous iron-rich chlorite. It is a green fibrous chlorite occurring in small pockets interstitial to the quartz, and has been determined as chamosite by x-ray diffraction.

The siltstone beds with a faint internal lamination display a wealth of depositional sedimentary structures. The upper portion of the individual siltstone bed commonly contains abundant cross-bedded scoops, each having a maximum depth of 4 inches and a length of about 12 inches.

There is also an unusual type of ripple mark (Plate 4-a) in which the internal laminations form symmetrical ripples with forward migrating crests. The wavelength averages 12 inches and the amplitude

2 inches. There is no truncation of the fore-set laminae, and they are continuous over the crest and down the back slope. The laminae are however thicker on the front slope. This structure is attributed (Sanders 1963) to shifting of sand by traction to produce ripples while there is considerable fallout of sediment from suspension.

Trains of discrete cross-bedded lenses of sandstone occur on the one bedding plane within the black shale. These are small crescent-shaped bodies of sand, about 12 inches long and 3 inches high, separated from each other by a distance of about 2 feet. They are analogous to subaerial barchans, and thus may be termed sub-aqueous barchans. These structures have been produced experimentally by Allen (1964, p.103) on a smooth bed where there is insufficient sand to form a continuous sheet. McCulloch and Janda (1964) have observed similar structures in the bed of the Nome River (Alaska) on a portion of the bed having a continuous pebble cover. In the example at Cowrie Point there was a continuous mud cover. This is partly explained by Hjulstrom's Diagram (Hjulstrom, 1935). It is possible for a current velocity of say 20 cm per sec to pick up sand particles by traction yet leave the layer of mud unscoured. This is due to the lack of hydraulic forces concentrated on the small particles, and also because of the greater cohesion of mud.

Primary current lineation is absent from the siltstone, and flute marks at the base of graded siltstone layers are rare. Such layers, where present, commonly contain angular black mudstone fragments. This suggests erosion and deposition by bottom-hugging turbidity currents, however, it is clear that these currents played a minor role in the deposition of the Cowrie Siltstone.

Black River.

At Black River the Cowrie Siltstone is a well-bedded and finely laminated shale which varies in colour from purple to pale green (33279). The finer grained laminae contain over 70% matrix of fine sericite, very fine quartz chips, irregular shreds of red opaque material probably colloiddally precipitated hematite, chlorite and other unidentifiable clay minerals. The coarser laminae contain up to 70% of medium silt-sized quartz and clastic fragments of muscovite. The varvoid shale is interbedded with black, pyritic, non-laminated mudstone.

Mawbanna district

Over large areas in the Mawbanna district, the Cowrie Siltstone is a finely laminated, (33250, 33251, 33252, 33253) silty shale. It is characterized by a conspicuous colour banding due to fine alternations of light gray claystone and brown fine-grained siltstone. The laminae vary from a fraction of a mm up to 10 mm,

and even the paper-thin layers show clearly as a continuous banding. The finer grained layers (33253) are composed of a mixture of very fine sericite and small chips of quartz up to 0.025mm. There are abundant vermicular and needle-like aggregates of barely resolvable birefringent mineral, possibly rutile. The siltstone layers are medium to fine-grained with equal proportions of sericite and quartz grains averaging 0.01mm. The colour is due to limonite staining of the micaceous minerals and also to irregular trails of finely divided graphitic material. There is no grading in the layers and exotic pebbles are absent. Many of the silty layers contain a fine internal lamination and wavy cross-bedding. Iron oxide concretions up to 5cm in diameter occur in these rocks. These are disc-shaped bodies and are characterized by the undisturbed passage of the lamination through the concretion.

Sisters Hills

In the eastern part of its outcrop, the Cowrie Siltstone is coarser grained and loses much of its distinctive fine lamination. Specimen 33261, from the Detention River where it flows through the Sisters Hills, is a laminated siltstone consisting of thin alternations of coarse siltstone with 10% sericitic matrix, and medium siltstone with 70% matrix.

Specimens 33255, 33256 and 33257 from the railway cutting through the Sisters Hills are typical of the more coarsely grained and more thickly bedded siltstones close to the top of the formation. These consist of poorly sorted angular quartz fragments (0.01mm), about 5% of randomly oriented clastic mica (0.5mm), and up to 40% matrix of fine crystalline sericite with fine chips of quartz. Detrital tourmaline, zircon and magnetite are accessories.

An increase in grain size of the siltstone and an increase in thickness of bedding units marks the passage into the overlying orthoquartzite of the Detention Sub-group. This is accompanied by a decrease in the amount of matrix and a decrease in the proportion of the lutite. The transitional rock types (33260) consist of poorly sorted, quartzose sandstones with 10 - 15% matrix, alternating with coarse siltstone. Bedding units range from 2 inches to 2 feet in thickness. These rocks contain abundant penecontemporaneous deformational structures which are described in Chapter 5. Ripple marks are common in the siltstone.

DETENTION AND JACOB QUARTZITES

Overlying the Cowrie Siltstone is the Detention Sub-group which is followed by the Irby Siltstone and the Jacob Quartzite. The two orthoquartzite formations are similar in lithology and are discussed together.

The Jacob Quartzite differs from the quartzites of the Detention Sub-group in its finer grain-size, thinner cross-bedding units, more consistent cross-bedding direction, the greater abundance of ripples and more shaly layers.

In its typical aspect, the orthoquartzite is a well-sorted, white quartz sandstone characterized by abundant cross-bedding. The bedding units vary in thickness from 2 inches to 6 feet and contain faint internal laminations. Interbedded conglomerate is absent.

In hand specimen the orthoquartzites are uniformly fine-grained sandstones with a texture varying from saccharoidal to glassy. Some of the coarser sandstones have a granular texture. In thin section they are seen to consist of over 97% quartz in the form of original grains and quartz cement.

Specimen 33264 is from the top of the Detention Sub-group on the Bass Highway on the western flank of the Sisters Hills. It is composed entirely of medium to fine sand-size quartz grains with a silica cement. They average 0.2mm in diameter, are well sorted, and generally equant in shape. The original shape is mostly modified by condensation of the fabric, for the grains are tightly packed with no void space. All grain-to-grain contacts are either planar, concave-convex or sutured.

Tangential contacts are absent. Many of the grains show undulose extinction. The original shape, where preserved by the filling of void space with silica cement, is well rounded and spherical. Both matrix and detrital muscovite are absent from this specimen, and there is 1% of well rounded tourmaline.

Specimen 33266, from the eastern flank of the Sisters Hills on the Bass Highway is similar to 33264, but contains 3% of chert grains. These fragments occur either as rounded grains, or as squashed lumps, interstitial to the quartz grains.

Specimen 33265, from the crest of the Sisters Hills, is not so pure a quartz sandstone, but has orthoquartzitic affinities. It consists of 80% of well-rounded, coarse sand-size quartz grains with overgrowths. Other components include 8% of slightly sericitized orthoclase, 5% rounded chert, 2% microcline, and 5% sericite matrix in isolated lumps.

Specimen 6111 from the old port at Rocky Cape (Cave Quartzite Formation, of the Detention Sub-group) is virtually 100% quartz. The texture is variable even in the field of one thin section. There are irregular patches where the unmodified detrital grains are easily recognised by the quartz cement overgrowths and the fine dusty trail on the inner edge of the overgrowth. Original porosity has been reduced to zero by complete cementation

of void space. In these zones the grains show tangential or planar contacts, or more rarely planar or slightly curved contacts due to pressure solution. Some idea of the rounding and sphericity can be obtained by visual assessment of the unmodified patches. Estimates from a visual comparator (Rittenhouse 1943), show the sphericity is about 0.8 to 0.9. The grain outlines form broad curves with no re-entrant angles. Sorting appears to be good, with the diameters ranging from 0.3 mm to 1.2 mm, which is only three intervals on the Wentworth grade scale. The greater portion of the grains are in the medium-sand interval.

In patches of greater textural modification, the texture is a closely interlocking mosaic with concave-convex, or more commonly, sutured grain boundaries, (Plate 2-a). The grains are elongate with their longest dimension parallel to the bedding. This is probably a pressure solution effect due to loading. Where overgrowths can be recognised, it is seen that suturing affects both the authigenic quartz and the detrital grain. This suggests that cementation occurred before suturing, and that the cement is not the material liberated locally by pressure solution of grains.

Specimen 33289 is from one of the minor gritty beds in the Jacob Quartzite, one mile west of Jacobs Boat Harbour. This is not so well sorted varying from

0.2mm to 4mm and taking in five intervals on the Wentworth scale. It is composed of 98% quartz with some rounded fragments of microcrystalline quartzite and sericite in small interstitial pockets. The original grains are clearly visible by virtue of well developed overgrowths (Plate 2-b). The grains are well rounded and have a high sphericity. The grains show tangential contacts, and all void space is reduced by cementation. This also suggests that cementation occurred before compaction.

Interbedded lutite in the orthoquartzites

Fine-grained siltstone beds comprise about 10% of the orthoquartzite formations. The lutite beds vary in thickness from a few feet up to 150 feet. They are of limited lateral extent, and appear to pass into the orthoquartzite by inter-digitation rather than by simple lensing.

The thickest of these horizons is the Port Slate in the Detention Sub-group, which occurs on the western side of Rocky Cape. The bottom contact is gradational, the underlying thickly bedded orthoquartzite (Bluff Quartzite) passes upward into a flaggy quartzite interbedded with thin layers of grey siltstone. The quartzite layers are cross-bedded, and as they thin to about 2 or 3 inches, they form discrete cross-bedded lenses.

The bulk of the Port Slate consists of abundant cross-bedded sandstone lenses in a grey siltstone (Plate 3-a). Some thicker beds of sandstone can be followed continuously across the shore platform for up to 50 yards.

These finer grained beds acted as incompetent horizons during folding and in places contain abundant drag folds, strong slaty cleavage (33267) and oblique shear faults. Where cross-bedded lenses are present, the rock superficially resembles a stretched pebble conglomerate, and had deformation been accompanied by recrystallization they may have become indistinguishable from sheared conglomerates.

The interbedded lutites contain some unusual structures which have been noted by Spry (1957a p.83) and likened to worm-tracks. These problematical structures occur in the small road-side quarry four miles from Sisters Beach, in the Port Slate, and at Rocky Cape are abundant in the fault wedges of Irby Siltstone about $\frac{1}{2}$ mile east along the coast-line from the old port at Rocky Cape.

These structures occur in a finely laminated, clean sandstone and claystone, (33290). The claystone laminae are broken and twisted, and on fractured bedding surfaces they appear as tubicolar threads. When the rock is cut across the bedding all gradations between regular lamination and completely disrupted laminae can

be seen (Plate 3-b). Even when strongly developed, the gross-aspect of bedding is still preserved so that an origin by lateral slumping is unlikely. They are unlike mud-cracks. They are probably the result of soft sediment load adjustment initiated and accentuated by small-scale lensing of the sandstone laminae.

Cross-bedding and related structures in the orthoquartzite

The cross-bedding is mostly the Omikron type of Allen (1963b). It occurs as large-scale, grouped sets, each set underlain by essentially planar erosional surfaces. Each set is broadly tabular or acutely wedge shaped, tapering back along the current direction, (Plate 4-b). The cross-bedding occasionally occurs in solitary sets, and more rarely, in cosets each with a concave basal plane, (Plate 5-a).

The cross-beds have a slight curvature, although the basal contact is not quite tangential. The upper contacts are sharp, and show no tendency to become tangential. In the plane of the bedding, the traces of cross-bedding on planar bedding surfaces are mostly straight, or less commonly, gently curved. The cross-beds are defined by a faint lamination due to slight grain size variations. The data for the thickness of individual cross-bedded units are summarised in Figure 3.

FREQUENCY DISTRIBUTION OF THICKNESS OF CROSS-BEDDED SETS

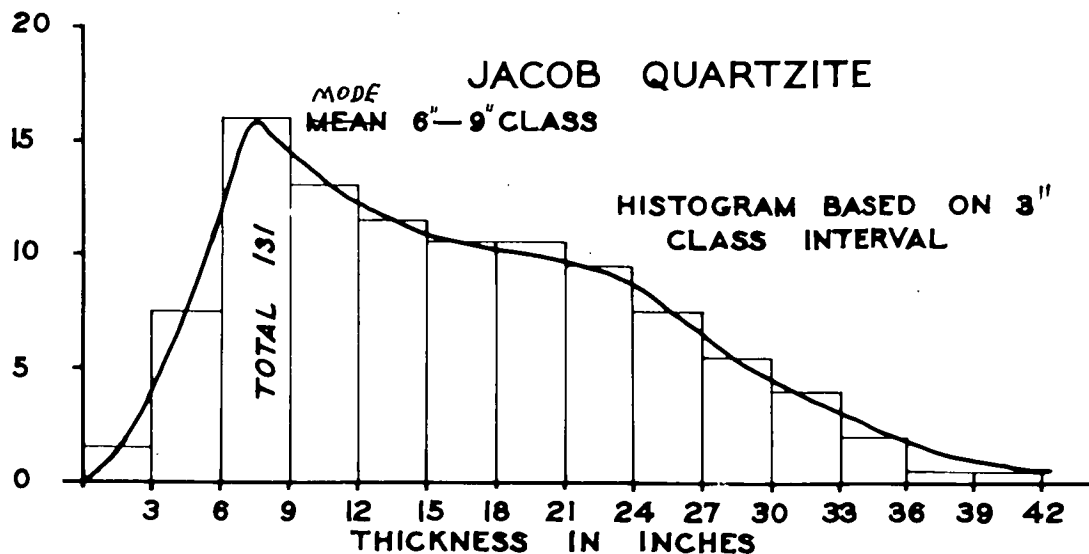
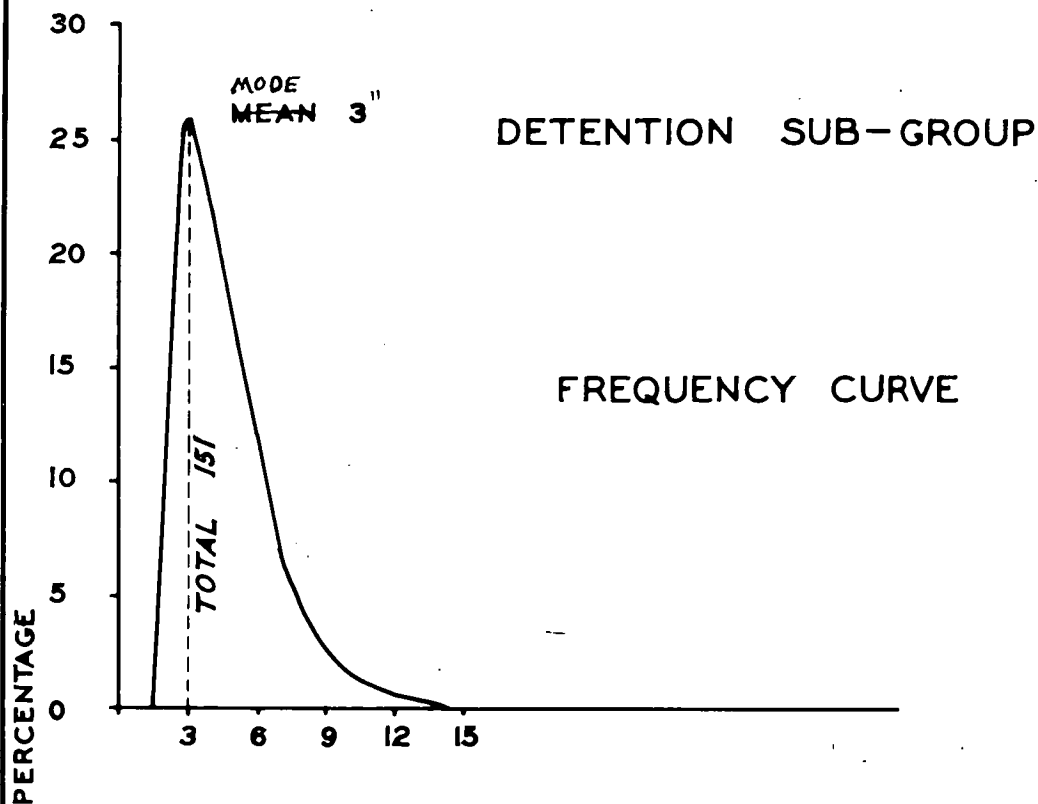


Figure 3.

Dragged-over cross-bedding has been observed (Plate 5-b) in the Jacob Quartzite where the cross-bedded units are thicker. This phenomenon has been experimentally investigated by McKee (1962) who considers it due to drag by a swiftly moving current. In the three examples observed, the axis of rotation was perpendicular to the cross-bedding dip, indicating a fixed relationship between the deformation and the current direction.

Azimuths of cross-bedding

An analysis of cross-bedding has been made to determine the patterns of the palaeocurrent directions. Field measurements that have been recorded include the dip and strike of cross-bedding, the dip and strike of the normal bedding, the trend and plunge of the local tectonic axis, and the thickness of the cross-bedded unit.

The regional folding is not considered to unduly complicate the determination of the tilt-corrected current directions. The method described by Norman (1960) for unrolling sedimentary lineations on plunging cylindroidal flexural folds is followed. The palaeocurrent direction is the line lying in the bedding plane, perpendicular to the line of intersection of cross-bedding and the normal bedding. The tilt-correction is made in two steps, firstly a plunge correction along the shortest

path to bring the tectonic axis to horizontal, and secondly a rotation about the fold axis to bring the bedding to horizontal.

The assumption here, is that the folds attained a plunge by a rotation about a horizontal axis at right angles to the fold axis. It is immaterial whether this results from two distinct phases or from two processes acting synchronously. Cummins (1964) has shown that this simple method is incorrect if the plunge is due to a rotation about an axis which is not at rightangles to the fold axis. The initial plunge correction is then not taken along the shortest possible path. This situation would be met with in repeatedly deformed areas, and a more refined and difficult method, taking into account the strain axes of each phase, is necessary.

The structure in the Rocky Cape Group is geometrically simple. Inland, the major and minor fold axes trend 060° and have a shallow plunge. Toward the coast the plunge steepens to 50° on the same general trend. The axial-planes remain vertical so that any rotation of the fold axes as they were formed must have been in the vertical plane. The additional plunge correction suggested by Cummins (1964) is not necessary.

Because the cross-beds have a slight curvature at the bottom, dips and strikes are measured as near as possible to the upper contact to give the maximum inclination. The angle of inclination is the dihedral angle between bedding and cross-bedding, and is given by the great-circle distance between respective poles in stereographic projection.

A total of 283 measurements of cross-bedding have been taken from 13 stations spread over 8 miles of coastal exposure. Axes of 44 ripple marks have also been measured. An areal coverage was not possible as inland outcrops are poor. A sampling procedure was devised after a preliminary sample of two stations revealed that with about five measurements a very strong pattern of preferred orientation emerged, but as more measurements were taken, other equally strong directions emerged. In view of the polymodal patterns the following procedure was used:-

- (a) Localities were selected along the fore-shore at about $\frac{1}{2}$ mile intervals, in an area where the bedding was regular and free from minor folds.
- (b) A station was selected at the locality, either arbitrarily or where there was abundant cross-bedding.

- (c) Commencing on any one cross-bed, and working stratigraphically down the section, every cross-bed was measured. Care was taken not to miss a set, and not to measure the same set twice.
- (d) At least 10, and generally 20, in one case 50 sets were measured. Thus between 20 and 50 feet of section is exhaustively examined. This sampling procedure collects many measurements from few stations so that the emphasis is on variability of current direction within a small section.

The tilt-corrected azimuths for each station have been grouped into 30° class intervals and plotted as rose diagrams in Figure 4. These give the direction toward which the main currents were flowing. The bimodal patterns and the large number of observations per station render this type of treatment more informative than vector summation, as is commonly done. For these diagrams, vector summation would be completely misleading. For example, the vector mean of station 4 is 290° , but there is only one of the 20 measurements lying in this sector. Vector modes may, however, be used to determine the mean direction for each mode. These are obtained by the vector addition of the azimuths which make up a single mode. For this purpose, sectors of 6° or less have been

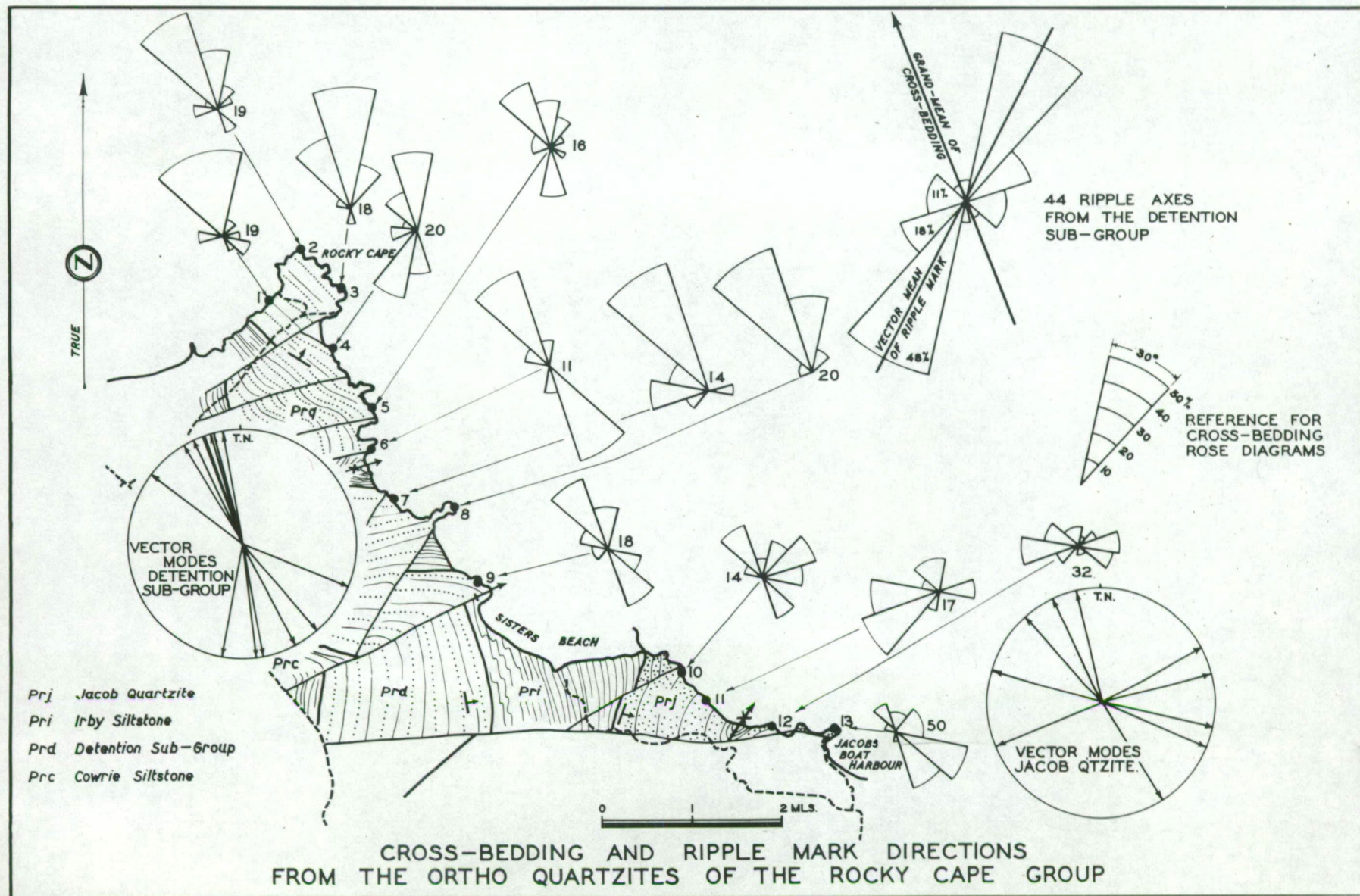


Figure 4.

neglected. The vectors are summed, as outlined by Reiche (1938), by assigning unit length to each azimuth. The ratio of the length of the resultant vector to the sum of the individual vectors (C/N) is called the consistency ratio and gives a measure of the dispersion of individual current components. The less the scattering of the azimuths, the closer C/N approaches unity. The vector modes and consistency ratio of each mode are tabulated in Table 1.

In general, there are two types of rose diagrams, namely unimodal and diametrically opposed bimodal. In the Jacob Quartzite there are also two trimodal diagrams. Azimuths constituting a single mode have a strong preferred orientation, shown by the high values of C/N . Azimuths constituting a single mode are also strongly segregated in the measured section. For example, the first fourteen measurements at station 12 constitute the easterly mode, and the following 18 constitute the westerly mode.

The mutual relations of the structure and the coast-line allows stations at different stratigraphic intervals at the same palinspastic position to be compared. Stations 13 and 1 show very similar patterns as also do stations 8 and 7. In the Detention Sub-group, the paleocurrent system was essentially constant for the duration of deposition.

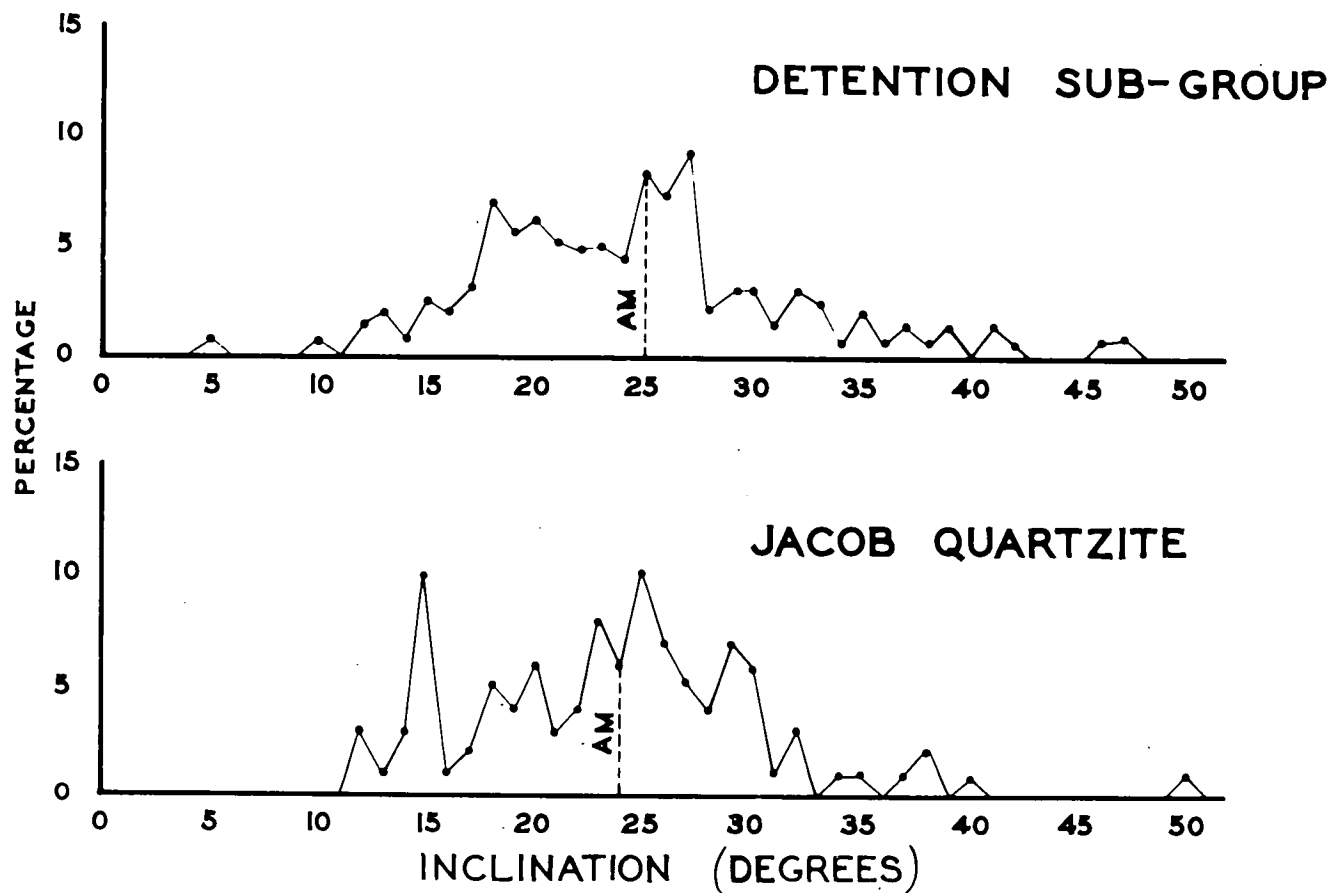
Rock unit	Station	Vector mode	R/n of mode	Description
JACOB QUARTZITE	13	100(degrees) 322	0.78 0.94	bimodal
	12	089 274	0.88 0.85	bimodal
	11	062 234 335	1.0 0.82 0.09	trimodal
	10	048 135 306	0.91 0.99 0.92	trimodal
DETENTION SUB-GROUP	9	156 317	0.97 0.98	bimodal
	8	329	0.88	unimodal
	7	295	0.67	unimodal
	6	127 322	0.96 0.93	bimodal
	5	159 331	0.97 0.87	bimodal
	4	177 334	0.98 0.96	bimodal
	3	139 332	0.98 0.95	bimodal
	2	339	0.90	unimodal
	1	099 330	0.95 0.92	bimodal

Table 1: Vector modes and consistency ratios of individual modes for all sample stations.

The regional variation of the azimuths and the vector modes are shown by the rose diagrams in Figure 4. In the Detention quartzite, most of the azimuths trend northwest, and the vector modes are closely grouped. The only significant variations are the presence or absence of the diametrically opposed southeast modes. In the Jacob Quartzite cross-bedding directions are more variable. The northwest modes are still present but not dominant, and the variability is due to the presence of modes to the southwest, southeast and northeast, Figure 4.

Inclination of cross-bedding

Figure 5 is a frequency distribution of the inclination of cross-bedding for both Jacob and Detention quartzites. The distribution is approximately normal, and the irregularity may be due to insufficient sampling. The arithmetic mean of inclination for the Detention and Jacob Quartzites is ^{25.0° and} 24.9° respectively, and these correspond closely to the mode. Of special interest is the large spread, and the 20% of inclinations above 30°. The critical angle of repose of sand in water is 30°, but varies slightly with grain size, sphericity and sorting. Inclinations above 30° require an explanation and three factors may be responsible.



**FREQUENCY DISTRIBUTION OF INCLINATION
OF CROSS-BEDDING**

Figure 5.

(a) Soft sediment deformation.

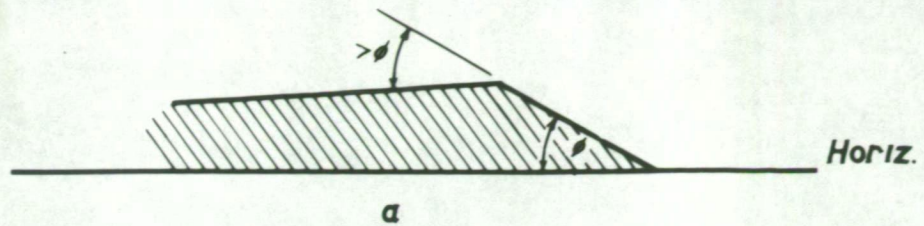
This takes the form of steepening, overturning or buckling of the cross-beds. This phenomenon is commonly recorded in the geological literature (Potter and Pettijohn 1963, p.80), and has undoubtedly been a factor in some cases, (Plate 5-b). The high inclinations of 40° - 50° are thought to be of this origin.

(b) Deformation during folding.

Folding will affect the angle of inclination. For fold types with a flattening component, the effect is difficult to evaluate (Ramsay, 1961), but for concentric deformation involving only bedding plane shear the problem is much simpler, and Brett (1955) Pettijohn (1957), and Pelletier (1958) have provided examples.

(c) Deposition on an incline.

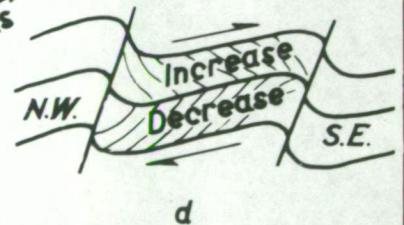
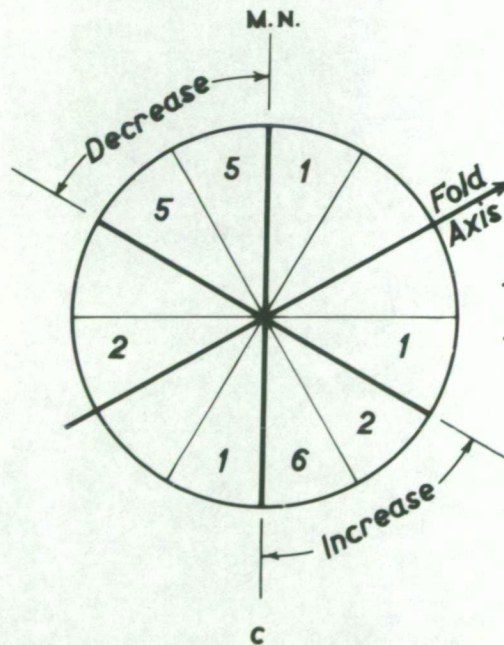
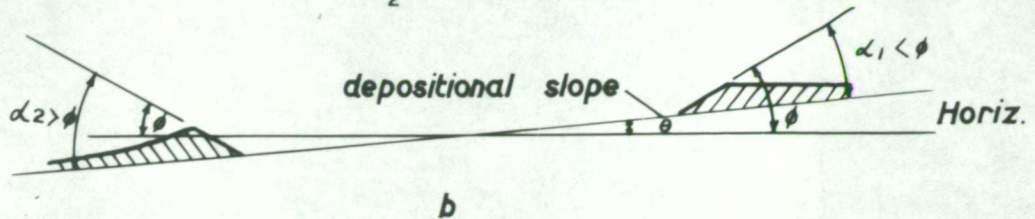
The inclination will either be larger or smaller than the true inclination depending on how the dihedral angle is measured. If the angle is measured between the cross-beds and the top truncating surface, the angle may be too large because the cross-bedded units lense in a direction opposite to the current direction, as shown in Figure 6. If, on the other hand, the bedding proper is used, whether it be above or below the cross-bedded set, the angle may be a fraction smaller or larger than the true angle, depending on whether the cross-beds face down or up the slope of



ϕ = Angle of repose

θ = Slope of depositional surface

$$= \frac{\alpha_2 - \alpha_1}{2}$$



- a Error in angle of inclination using top truncating plane.
- b Effect of deposition on an incline on angle of inclination.
- c Distribution of cross-bedding azimuths above 30° in Detention quartzite in relation to regional fold axis showing sectors where inclination should increase or decrease as shown in Fig. d .

Figure 6

deposition. In this study the second method was used so that any systematic variation may reflect the regional slope.

Factors (b) and (c) are difficult to distinguish because the tectonic axis which trends northeast-southwest is perpendicular to the main current direction. For ideal concentric deformation, the distortion of cross-bedding will be more fully recorded by those sets that have (restored) dip directions perpendicular to the fold axis. Cross-bedding sets that have dip directions parallel to the fold axis will appear to record no deformation.

In order to evaluate the effects of folding, the azimuthal distributions of inclinations above 30° in the Detention Sub-group are examined (Figure 6-c) in relation to the sense of shear expected from the overall structure shown in Figure 6-d. Also shown are the sectors, symmetrical about the fold axis, in which the inclinations should increase or decrease. The distribution is diametrically bimodal as would be expected from the overall pattern of azimuth distribution, and it shows more high inclinations in the decreasing sector than it does in the increasing sector. This is contrary to what would be expected if distortion were due to tectonic deformation. There can be no possibility of local reversals on minor folds causing local increases where there should be decreases because the structure is

sufficiently well known. Furthermore, there are some high inclinations with azimuths directed along the tectonic axis. It is therefore concluded that folding has had no recordable effect on the inclinations of the cross-beds. The structural implications of this are discussed later.

In Table 2 are compared the inclinations of various groups of stations in the Detention Sub-group, which are directed into the northwest and southeast sectors. It is seen that the inclinations of the northwesterly dipping cross-beds are consistently less than those of the southeasterly dipping cross-beds. Also shown is the average of all inclinations irrespective of direction, and these values lie between the two limits. The difference in average inclinations between cross-beds in these two sectors is probably due to deposition on a regional slope that had an overall dip to the northwest. The magnitude of this dip in the northwest direction is given by half the difference between the opposing inclinations (Figure 6-b). For the Detention Sub-group, the values for the three groups of stations are 1.7° , 1.3° , and 2.5° . The sediment distribution pattern in the Jacob Quartzite is diverse and the ideal relationship between a regional slope and cross-bedding inclination does not appear to hold, (Table 2).

Rock Unit	Group of Stations	Mean Inclination	Mean inclination in sector		number of readings in sector	
			120°-180°	300°-360°	120°-180°	300°-360°
Jacob Quartzite Detention Sub-group	12	22.3°	23.2°	21.7°	5	6
	10, 11	24.3	24.0	27.9	6	9
	9, 8, 7	24.1	27.5	23.1	6	29
	6, 5, 4	29.2	32.2	29.7	11	20
	3, 2, 1	22.1	27.5	22.4	6	38

Table 2 : Comparison on inclinations for various groups of stations which have cross-bedding directions in the northwest and southeast sectors

Ripple mark

Ripple marks are uncommon in the Jacob Quartzite but locally abundant in the Detention Sub-group. At these localised concentrations of ripple mark, cross-bedding is either uncommon or absent. The ripples are the transverse type occurring in rectilinear or bifurcating sets. Wavelengths vary from 1.5 to 2.0 inches, and the ripple index varies from 7 to 13. inches. Both asymmetrical and symmetrical types are present, and many of the symmetrical ripples have secondary crests.

Forty-four ripple axes have been measured from various localities along the coastal outcrop of the Detention Sub-group. These are recorded in Figure 4. Each measurement represents the average of a set of rectilinear ripples. No distinction has been made between the different types of ripple mark, because all gradations between the two types are found. However, it has been observed that asymmetrical ripples indicate opposed currents from northwest and southeast with about equal frequency.

The significant features of Figure 4 are the strong clustering of ripple axes to the northeast, and the high angle between the grand-mean of the cross-bedding and the mean ripple axis. The minimum direction of ripple axes corresponds to the grand-mean of the cross-bedding direction. Ideally, transverse ripple mark is perpendicular to the dominant cross-bedding direction, and this has been

demonstrated in marine sandstones by Brett (1955) and Hamblin (1958); and in fluvial sandstones by Pelletier (1958) and ~~Parrie (1961)~~. Potter and Pettijohn (1963, p.94-98) have discussed the significance of ripple orientations and conclude that ripple axes delineate the local depositional strike, and hence the trend of the strand line.

IRBY SILTSTONE

The Irby Siltstone is composed mainly of siltstone with lesser amounts of black shale, dolomite, quartzite and sub-graywacke. The base and top are not exposed. A traverse from west to east along Sisters Beach (Figure 19) crosses the following rock types:

- (a) black siltstone at, or near the base, with minor cross-bedded lenses of sandstone, at least 130 feet;
- (b) laminated and thickly bedded dolomite, at least 300 feet;
- (c) interbedded sandstone, siltstone and mudstone, at least 510 feet;
- (d) a gap in the outcrop allowing for 1000 feet of section, assuming a continuation of the structure;
- (e) sub-graywacke sandstone and mudstone, at least 100 feet;
- (f) cream argillite, about 200 feet;

(g) Jacob Quartzite on the eastern headland of the beach.

These units are bounded by faults, but the structural evidence indicates that they form a sequence which is younging to the east. Some of these rocks contain peculiar structures, and are discussed below.

Dolomite

The main part of the dolomite consists of buff-coloured dolomite units from 1 to 4 feet in thickness, with minor inter-bedded dolomitic mudstone from $\frac{1}{2}$ inch to 1 inch in thickness. The dolomite has a fine internal lamination which commonly shows small scale cross-bedding. Load casting between the dolomite and mudstone is also present.

In thin section (33296, Plate 6-a), the dolomite is composed of a structureless granular mosaic of anhedral dolomite of uniform (0.05mm) grain size. It is studded with detrital quartz (0.05mm), and minor amounts of detrital muscovite. In this specimen, the dolomite shows no tendency to form euhedral crystals.

The following is a chemical analysis of the massive dolomite (33296) from Sisters Beach.

SiO ₂	40.9%
Al ₂ O ₃	4.7
Fe ₂ O ₃	0.8
FeO	2.2
MnO	0.06
P ₂ O ₅	0.12
TiO ₂	0.16
CaO	14.6
MgO	11.5
K ₂ O	1.0
Na ₂ O	Nil
H ₂ O-	0.14
H ₂ O+	1.5
CO ₂	22.0
SO ₃	0.17
S	0.16

Analyst: H. K. Wellington.

This analysis reveals that the molecular ratio of MgO/CaO is 1.10 indicating that the carbonate present is close to pure dolomite mineral. The rock however contains about 40% of quartz of a detrital origin, and is not a pure dolomite rock compared to the Smithton Dolomite, (Hughes, 1957, p.282).

The lower part of the dolomite horizon is a fine alternation of dolomite and mudstone, and exhibits good examples of contorted sedimentary boudinage. On the shore platform, weathering causes the dolomite boudins

to stand out in relief, producing "hieroglyphic" markings, (Plate 6-b). The undeformed rock appears to have been a regularly laminated brown dolomite in layers 5mm - 1.5cms thick, and thinner black mudstone about 3mm thick. The mudstone laminae are rich in fine sericite with about 10% of fine detrital quartz averaging 0.02mm, about 10% anhedral dolomite grains, and about 3% detrital muscovite. The dolomite layers in thin section are petrographically the same as the more thickly bedded dolomite, although they contain irregular patches of recrystallized euhedral dolomite. The dolomite layers have a faint internal lamination which also shows fine cross-bedding.

The zone of laminated dolomite is about 30 feet thick, and no base is exposed. At the bottom of this zone the dolomite layers show various degrees of segmentation without disorientation. Where this segmentation is complete, the dolomite pieces are completely surrounded by mudstone, although the form of the original lamination is still apparent. The dolomite pieces invariably have rounded edges, and in this respect are distinct from the "inhomogeneity breccia" of Sander (1951) which are formed by the in situ rupture of brittle layers within a more plastic material. Such breccias contain sharply angular pieces that in places can be matched.

The boudins are markedly elongate in a northeasterly direction, giving a conspicuous coarse lineation approximately down the dip of the bedding. This lineation is also parallel to the regional fold axis (50° plunge to 060°), and to the axes of minor folds and related bedding-cleavage intersections in the nearby cleaved mudstones. The tectonic cleavage in the dolomite is expressed as sinuous and finely corrugated surfaces that are broadly planar. Corrugation is due to the fractures seeking out the nodes of the boudins (Plate 7-a).

The boudin bodies are crossed by a set of black, hair-line, tensional cracks (Plate 7-b), which are perpendicular to the bedding and parallel to the axis of elongation. These cracks are up to 0.5mm wide and filled with the black mudstone material containing many recrystallized euhedral dolomite rhombs. There has been a slight amount of movement along the fine cracks.

A fine flow-lamination is seen in thin section in the mudstone. This is generally parallel to bedding but bows into the nodes and encircles the rounded ends of the boudins. The sericite of the matrix and the detrital muscovite are oriented in the micro-foliation, and wrap around the detrital quartz. Where the hair-line cracks intersect the mudstone layers a micro-strain-slip cleavage forms.

Further up in the boudinage zone, the dolomite pieces become disoriented and all continuity of the original fine layering is lost. Many of the pieces are bodily rotated into an edge-wise position at right angles to the bedding, this being especially common adjacent to the more prominent sinuous movement cracks. In these edge-wise pieces, the hair-line cracks are still at right angles to the slabs, showing that rotation occurred after micro-fracturing.

At the top of the boudinage horizon, the dolomite bodies become completely disoriented and twisted, (Plate 6-b). However, the gross aspect of bedding is still undisturbed showing there has been no bulk deformation of the rock, only an in situ segmentation, accompanied by independent internal rotations.

Immediately overlying the most complex bed is a bed of dolomite 8 feet thick, with a few minor mudstone layers. Boudinage is absent, and the cleavage is manifest by a weak, planar, close-spaced jointing.

A sequence of development of these structures can be followed, starting with a plastic segmentation of dolomite layers and viscous flowage of mudstone, then micro-tension fracturing, followed by coarser fracturing and associated tilting and bending of the dolomite pieces. The rounded edges of the boudins, the micro-flow foliation in mudstone, and the twisting and rotation of dolomite bodies suggest these are preconsolidation

structures. Recrystallization of the dolomite was not an important process in boudinage formation.

The obvious cause of boudinage is stretching in the bedding plane. Shrinkage during dolomitization of a limestone is ruled out because the fine granular anhedral texture, together with the fine layering and the fine cross-bedding suggest this is a primary dolomite. Also, such a process would be expected to yield a "chocolate tablet" structure rather than elongate structure. For this reason other processes of uniform shrinkage in the bedding plane during diagenesis, as envisaged by McCrossan (1958) and Greenwood (1960), are not applicable.

Boudinage formation generally, can be explained following Ramberg (1955) by plastic deformation of a layered material under a triaxial stress field oriented with the maximum principal stress normal to the layering, the intermediate stress parallel to the boudin elongation and the minimum stress in the plane of the layering and normal to the boudin axis. It is therefore possible that these boudins result from a directional shrinkage in the bedding plane which may be due to slight preconsolidation movement down the local palaeoslope. The regional palaeoslope, determined from cross-bedding measurements in the enclosing orthoquartzites, is probably to the northwest, and is in the required direction to give a northeasterly boudin axis. Nothing is known

of the local palaeoslope for the deposition of the Irby Siltstone.

The boudin orientation is related geometrically to the cleavage and the regional folding, with the boudin axis along the fold axis, although this is probably fortuitous. In so far as the fold axis may approximate to the major basin elongation, this configuration is not coincidental.

Interbedded sandstone, siltstone and mudstone.

The overlying unit consists of at least 500 feet and possibly 1000 feet of an alternation of greenish-gray siltstone and black mudstone, with minor quantities of cross-bedded orthoquartzite. The siltstone is coarsely laminated due to irregular ribbons of coarser grained siltstone, and contains an abundance of small-scale depositional sedimentary structures such as wavy cross-bedding, scour-and-fill structures, cross-bedded lenses, and irregular erosional planes. Bedding is further complicated by load casting. Specimen 33298 is typical of the siltstone. It is composed of 50% detrital quartz of medium silt-size, 15% finely splintered detrital muscovite, with magnetite, ilmenite, rutile, and tourmaline accessories. The matrix is fine sericite and chlorite.

The interbedded orthoquartzite occurs as single bedding units or an unbroken series of several units. They are petrologically similar to the thick orthoquartzites.

Sub-graywacke

The sub-graywacke unit near the eastern end of Sisters Beach consists of thickly bedded sub-graywacke, with thin alternations of fine siltstone. The sub-graywacke, varies in size from coarse silt to fine sand. It is green when fresh and weathers to a deep reddish-brown. Specimen 33302, typical of the coarse siltstone, consists of about 20% matrix of iron-stained sericite and chlorite, and abundant small hematite grains. The detrital grains are mainly quartz, with 5% feldspar (albite) and muscovite.

The finer grained siltstone is composed of sericite and chlorite with small quantities of detrital quartz. It is laminated with a glossy sheen on the bedding surfaces, and a weak oblique cleavage, and may thus be termed argillite. The argillite beds become thicker and more frequent at higher levels, and become the dominant rock type on the eastern headland of Sisters Beach.

Associated with the coarser sub-graywacke beds are some thin interbedded hematite breccias, in beds up to 12 inches thick. These occur between the thinly laminated argillite below and the massive arenite at the top (Plate 8-a). In this example the base is irregular due to scouring, whereas the top is smooth and conformable to the bedding. The breccia is composed of

thin flakes of solid hematite 1 - 2 mm thick and up to 20mm in length. Most are oriented flat in the bedding, but there is no evidence of continuity of bedding lamination which would be expected from in situ fragmentation. The flakes are occasionally twisted and crumpled. The matrix is a fine quartz siltstone heavily stained with iron-oxide. The smaller hematite flakes are lozenge-shaped and are wrapped by a fine streakly flow-foliation in the matrix (Plate 8-b). These features suggest open-cast lateral laminar movement in a high density mud-flow. The breccias may be termed fluxo-turbidites in the sense of Dzulynski, Ksiazkiewicz and Keen (1959).

SMITHTON DOLOMITE AND ASSOCIATED ROCKS

The Smithton Dolomite is a thick transgressive sheet of dolomite with a basal conglomerate in the Black River area. Most of the dolomite lies outside the area studied. This association is not included in the Rocky Cape Group, but is described in this chapter for convenience.

Forest Conglomerate and Quartzite

The Forest Conglomerate and Quartzite is about 50 feet thick and retains its thickness over two miles of the area investigated. It overlies the Cowrie Siltstone with angular unconformity. The details of this unconformity have been described in Chapter 2.

The conglomerate possesses a crude stratification due to a rudimentary sorting of the boulders. Where discernible, the bedding units are up to four feet thick. The conglomerate contains lenses of up to 12 inches thick of pure, white, well-sorted quartz-sandstone with a distinct granular texture in hand specimen.

The pebbles consist almost entirely of a well-sorted, pure, quartz-sandstone with a siliceous cement, identical in lithology to the orthoquartzite in the underlying Rocky Cape Group. Some rare fragments of siltstone, similar in lithology to the Cowrie Siltstone, are present. The fragments vary in size from $\frac{1}{2}$ inch up to 2 feet and the mean-size is about four inches. The fragments are all well-rounded and generally spherical. No imbricate structure has been observed.

The pebbles and boulders have a closed framework, the voids being filled with a matrix of pure quartz-sandstone composed of well-rounded quartz grains of medium to coarse sand grade. The immediate basal unit of the conglomerate is variable with locality. On the western abutment of the Bass Highway bridge over the Black River (Figure 17), it is a sand lens with coarse cross-bedding. In the railway cutting on the northern limb of the syncline it is a cobble conglomerate with cobbles up to three inches. In the Black River, one

and a half miles upstream from the bridge it contains boulders up to two feet in diameter.

The conglomerate passes upward into 20 feet of well-bedded clean orthoquartzite, which is commonly cross-bedded. Overlying the orthoquartzite is a thin horizon of white, finely laminated chert.

Smithton Dolomite

The dolomite has not been examined outside the Black River area. It is known to outcrop extensively in the area from Black River west to Montagu and Redpa. It has been described by Nye, Finucane and Blake (1934), Carey and Scott (1952), and Hosking and Hueber (1954). It is described as a light-grey to cream dolomite with thickly bedded coarse-grained varieties and fine-grained thinly bedded varieties. It contains well bedded cherts, and oolitic dolomite has also been reported.

At Black River it occurs in the trough of the syncline (Figure 17), outcropping sparsely in the river banks near the old highway bridge. Here it is grey to buff coloured, aphanitic and structureless apart from a diffuse stratification. It contains irregular lenticular nodules of chert which are elongate in the bedding plane. In thin section it is composed of about 95% dolomite grains (0.02 - 0.1mm) forming a granular mosaic, with irregular patches of recrystallized coarse, clear subhedral grains (0.2 - 0.4mm). Surrounding

the recrystallized patches is usually a film of sub-opaque, brown earthy material which forms somewhat regular shapes vaguely resembling organic remains. These are insoluble in acetic acid. Other insoluble residues include authigenic quartz and detrital muscovite.



Plate 1a Unconformity between Cowrie
Siltstone and Forest Conglomerate
showing ridges on the sole of the
bottom conglomerate bed.

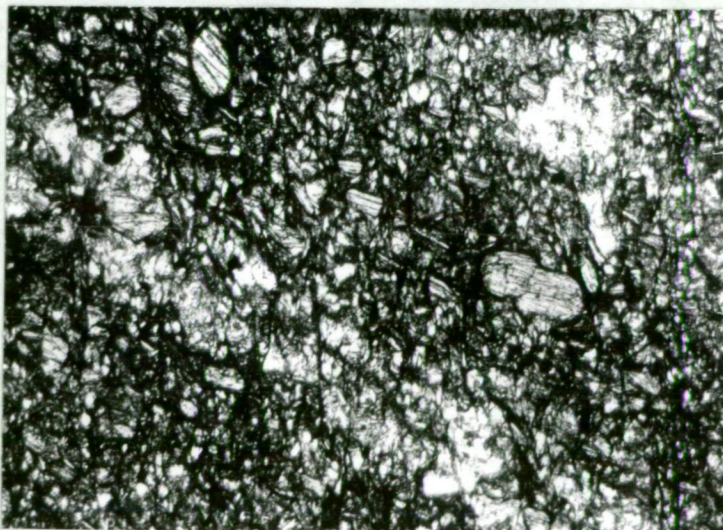


Plate 1b Photomicrograph of Cowrie Siltstone
showing rounded detrital flakes,
west side of Rocky Cape. Sp. 33273
x 70.

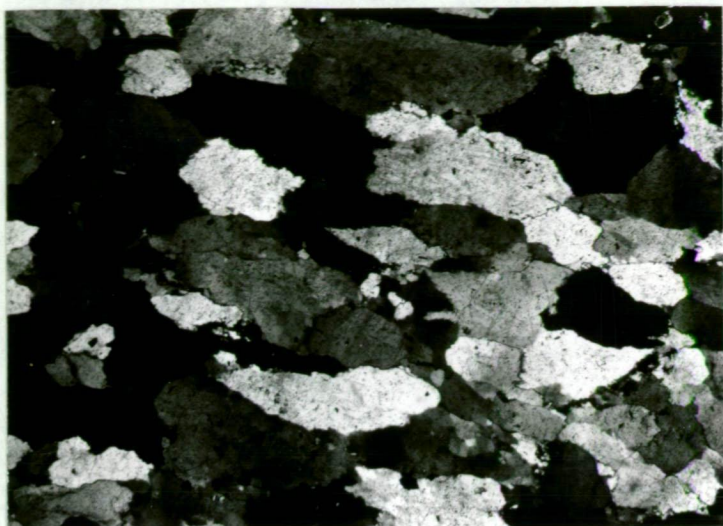


Plate 2a Photomicrograph of orthoquartzite
from the Detention Sub-group,
showing sutured grain boundaries.
Sp. 6111, x 22.

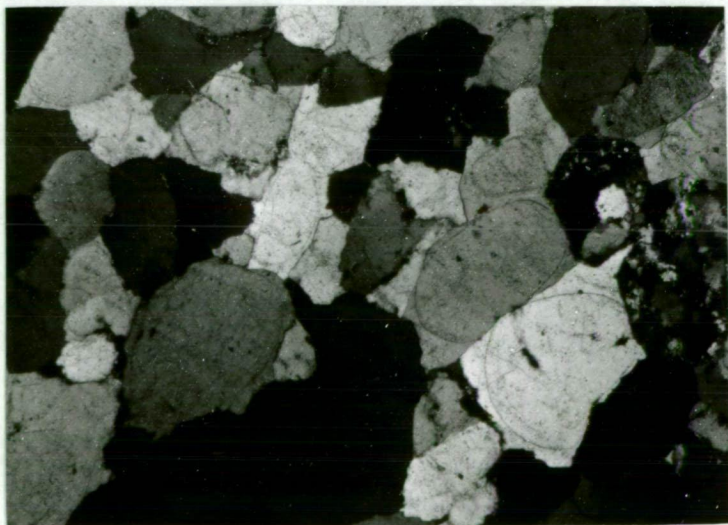


Plate 2b Photomicrograph of gritty bed in
Jacob Quartzite, showing over-
growths on rounded quartz grains.
Sp. 33289, x 28.



Plate 3a Cross-bedded sandstone lenses in
siltstone, Port Slate, Rocky Cape.



Plate 3b Disrupted sandstone laminae in
Irby Siltstone, $\frac{1}{2}$ mile east of
The Port, Rocky Cape. Sp. 33292,
x 2.



Plate 4a Asymmetrical ripples, Cowrie Siltstone, Crayfish Creek.



Plate 4b Cross-bedding in Jacob Quartzite, west of Sisters Beach.



Plate 5a Cross-bedding in cosets, Jacob
Quartzite, west of Sisters Beach.



Plate 5b Recumbent cross-bedding, Jacob
Quartzite, Jacob Boat Harbour.

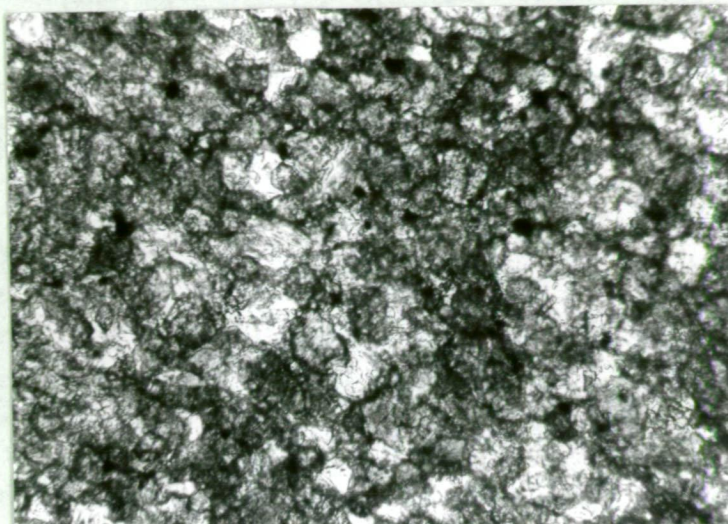


Plate 6a Photomicrograph of massive dolomite,
Sisters Beach. Sp. 33296, x 150.



Plate 6b "Hieroglyphic" markings in dolomitic
siltstone, Sisters Beach.



Plate 7a Sedimentary boudinage in dolomite
laminae in siltstone, Sisters
Beach. Sp. 33297, x 1.

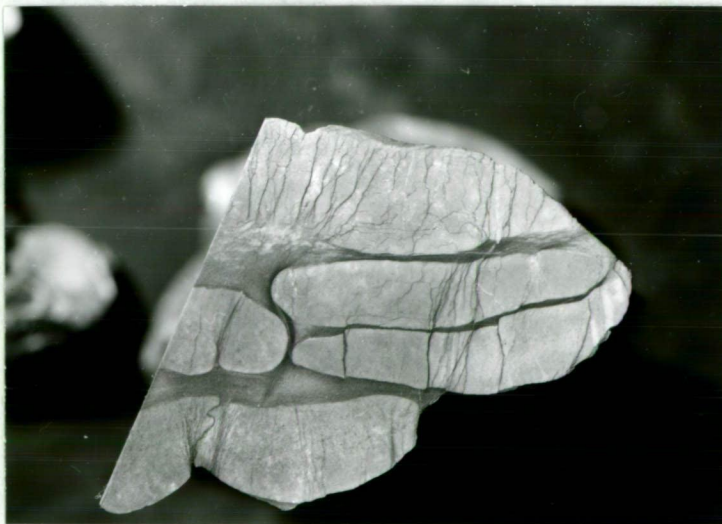


Plate 7b Dolomite boudins showing black
hair-line cracks. Sp. 33295,
x 2.

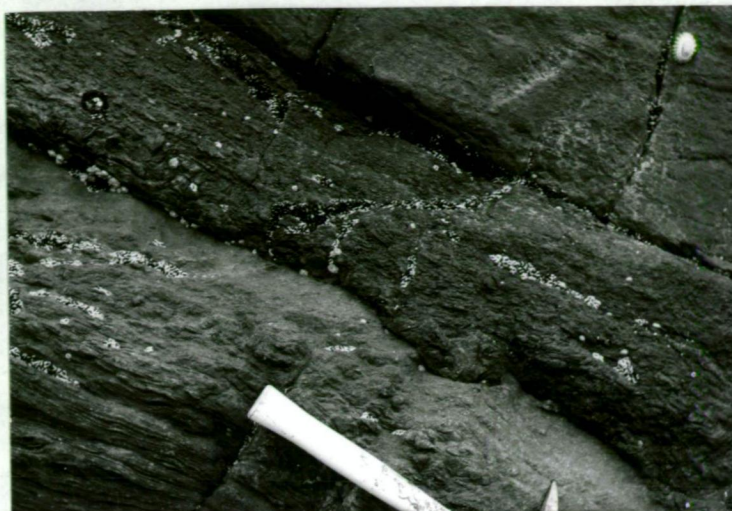


Plate 8a Fluxo-turbidite deposit of hematite breccia, showing scoured base and conformable top, top of Irby Siltstone, eastern edge of Sisters Beach.



Plate 8b Detail of hematite breccia, Plate 8a (above). Sp. 33301, x 4.

C H A P T E R 4

SEDIMENTARY FEATURES OF THE BURNIE FORMATION

The Burnie formation consists of at least 15,000 feet of a monotonous alternation of well-bedded black slaty mudstone, with quartz wacke of siltstone and sandstone grade. It is essentially uniform in lithology from Doctors Rocks to Sulphur Creek. The rocks at Cemetery Hill, Penguin, whose relative stratigraphic position within the Burnie Formation is uncertain, are better sorted than the usual arenite. These may be termed protoquartzite. Minor pillow lavas occur on the foreshore at Sulphur Creek, and possible pillow lavas occur two miles east of Doctors Rocks.

The following petrographic observations have been made by microscopic examination of 20 thin sections. Only four modal analyses of the arenites have been made, but thin-section examination shows that the arenites are remarkably uniform in composition and texture. Granular disaggregation of these rocks for the purpose of grain-size analysis is not possible, and purely qualitative categories of roundness, sphericity and sorting, based on visual estimates, are made. For estimation of matrix, an upper limit of 0.02mm for quartz grains is used following Pettijohn (1957, p.284).

PETROGRAPHY

Arenite

The arenites are tough rocks, gray to cream in colour when fresh and weathering brown. The rocks of coarse-silt and fine-sand size are featureless in hand specimen and the granular texture is not obvious because of the matrix coating on all detrital grains. In the more coarse-grained arenite at Howth, the larger detrital quartz grains can be recognised as black, angular fragments.

Mineralogically, the rocks are simple. Quartz is by far the most important clastic component. It occurs almost invariably in single crystallographic units. Both strained and unstrained quartz are present and several factors suggest that the clastic grains were already strained before incorporation into the rock. Thus, strained grains occur immediately alongside unstrained grains, and specimens lacking cleavage contain strained quartz. Because of the high proportion of matrix, it is improbable that any strain effects could be produced through compaction. In fact, the textural evidence of the cleaved arenite indicates (Chapter 11) that even the deformation has had little effect on the detrital quartz grains.

Many of the larger grains show clear evidence of recycling in the form of corroded and abraded quartz overgrowths (Plate 9-a, 33340). These are especially

common between Howth and Sulphur Creek. Parallel trails of fluid inclusions are common. Other inclusions in quartz are zircon, rutile, tourmaline, sericite and vermicular chlorite.

Feldspar has not been observed in any part of the Burnie Formation.

Muscovite is invariably found throughout, but rarely exceeds 10%. It occurs mainly in small slender flakes, often splintered by the imposed tectonic cleavage. More rarely it occurs in books of rounded outline. Minor amounts of detrital biotite and chlorite are present in the coarser arenite at Sulphur Creek.

Rock fragments include chert, fine-grained quartzite, clay aggregates all of similar grain size to the detrital quartz, and larger mudstone flakes. Chert is a minor constituent in the normal arenite, but reaches up to 15% in the quartzite at Penguin, where it occurs as rounded grains or squashed aggregates interstitial to the quartz grains, (33352, 33354, Plate 9-b). The chert is in the form of a micro-crystalline mosaic of quartz grains, (0.008 - 0.02mm). The fine-grained quartzite fragments which have the same mosaic texture may be due to a progressive coarsening with time of the mosaic due to recrystallization.

The clay aggregates are composed of fine sericite with some chlorite. These are of comparable size to the

surrounding quartz grains and thus appear also to be detrital. Most are rounded aggregates with smooth outlines, but a few are squashed between detrital quartz grains to form an interstitial paste, indistinguishable from the matrix material. Allen (1962) has also observed this feature, and suggests that much of the "graywacke" texture of the Old Red Sandstone results from the compaction of a mixture of clay aggregates and quartz grains. This does not appear to be entirely true in this case, because when unmodified, the initial shape of the clay aggregates is apparent, and a true interstitial matrix (often slightly more granular), is recognisable.

The origin of the clay aggregates is uncertain. There is no evidence that they are altered detrital feldspar, and feldspar has not been observed in these rocks. Flocculation of suspended clay material on entering sea-water immediately prior to deposition is a possibility. This process occurs by absorption of electro-positive ions into mutually repulsive, negatively charged clay particles. However, experimentally produced flocculated aggregates are usually of the order of 10-100 microns which is much smaller than the clay aggregates in question. It appears that the aggregates are detrital claystone particles that were transported and deposited along with the quartz grains and matrix.

The angular, wispy flakes of black mudstone are locally common, especially near the base of graded arenite beds. These are composed entirely of aggregates of minute crystallites of sericite showing aggregate polarization in the direction of the tectonic cleavage.

Accessory detrital minerals include rounded tourmaline, magnetite, hematite, sphene, zircon and rutile.

Matrix consists of fine crystalline sericite, green fibrous and vermicular chlorite, fine quartz chips (less than 0.02mm), gray semi-opaque earthy clay minerals and small splintered clastic muscovite. Much of the matrix is stained brown by limonite. Shreds of black graphite material are common. Small cubes of pyrite in the matrix are probably of authigenic origin.

The borders of the quartz grains are commonly hazy and diffuse due to penetration of chlorite and sericite. This diagenetic encroachment of the matrix over the quartz modifies the original degree of roundness of the grains, but probably does not greatly alter the shape (Plate 9-a). There are no deep embayments.

Secondary cement is not a feature of the arenite, and only one case has been found, (33349). This contains a carbonate cement that has replaced part of the matrix of a coarse arenite. This occurs close to the pillow lavas at Sulphur Creek, and it is possible that the

carbonate is diagenetically derived from deuteritic leaching of the sub-aqueous lava flows. Silica cement has not been found,

. Texturally, the coarse arenites possess a typical graywacke texture (33340, Plate 9-a) with a disrupted framework of poorly sorted quartz detritus surrounded by matrix, and with randomly oriented detrital muscovite. The grain size varies from a maximum of 1.5mm down to exceedingly fine chips in the matrix. The mean grain size (frequency mode of 300 diameters in thin section of grains over 0.02mm) is 0.35mm.

The larger grains in Plate 9-a have secondary quartz overgrowths in optical continuity. The original grains, defined by the border of dusty inclusions, are well rounded with a high sphericity. Diagenetic replacement has destroyed much of the grain boundaries and partly removed the overgrowths. The grains with secondary enlargements are thus recycled and may have been derived from the orthoquartzite formations in the Rocky Cape Group to the west, which are composed entirely of well-rounded, and spherical grains commonly with secondary overgrowths. The smaller grains in the coarse arenite at Sulphur Creek are generally angular, and many are seen to be fragments of initially rounded grains.

The finer grained arenites have a grain size in the medium to coarse-silt range. These consist (33351,

33319, 33329, 33331) of angular quartz grains with rare fragments of a once larger rounded grain. Clay aggregates also occur. These rocks do not display such a good disrupted framework as the coarser arenites, and some sutured contacts between grains are found. Plate 9-b, (33329) is typical of the texture of these rocks.

Corrosion by the matrix makes difficult the distinction between clasts and matrix. Splintered detrital muscovite in flakes up to 0.5mm in length, occur in random orientation.

The arenites appear to have no significant compositional variation over the major part of the Burnie Formation. Specimen 33340 from Howth contains 48% quartz, 3% detrital muscovite, 1% clay aggregates and chert fragments and 48% matrix. Specimen 33351 from Sulphur Creek headland contains 54% quartz, 3% detrital muscovite, 1% chert and 42% matrix. Specimen 33320 from Cooe Point contains 60% quartz, 3% detrital muscovite, 1% accessories and 36% matrix.

Terminology and classification of the arenite

The arenite in the Burnie Formation has a uniform lithology over the major part of this thick sedimentary basin and clearly represent a natural class of sedimentary rock of distinct lithogenetic significance. However, they are difficult to classify according to some of the more widely used classifications. They are

characterised by abundant matrix, so that classifications which neglect matrix as a parameter are not useful. Thus, according to Krynine (1948) and Folk (1954) many would plot in the quartz-feldspar-rock fragments ternary diagram as orthoquartzites.

Pettijohn's classification (1957, p.291), although using the matrix as a parameter, makes no provision for rocks which contain more than 15% matrix, and a high detrital quartz component. This arises because of the apparent contradiction of a rock having mineralogical maturity and textural immaturity. The arenites in the Burnie Formation contain up to 45% matrix and should be grouped with the graywackes. Most are clearly not sub-graywackes in Pettijohn's sense (p.301) because they contain more than the permissible 15% of matrix.

This large group of rocks has been recognised by Gilbert (Williams, Turner and Gilbert 1954) and later by Reed (1957) and Dott (1964). Gilbert defined a class of sandstones composed of more than 10% matrix as wackes. These may be further subdivided if they contain unstable detrital grains (lithic wacke and feldspathic wacke), or stable constituents (quartz wacke). Gilbert also proposed a narrow use of the word "arenite" for well sorted and mature sediments containing less than 10% matrix, so that arenites and wackes were the two basic types of sandstones.

Although this step is open to question since it removes the last all-embracing term for sediments having a certain grain size, it does not detract from the value of the word "wacke".

The arenite in the Burnie Formation is therefore termed quartz wacke.

Interbedded lutite

The lutite interbedded with the quartz wacke varies from black, featureless, or weakly laminated mudstone, through dark grey argillite, to gray glossy slate in which the cleavage is oblique to bedding. It is mineralogically similar to the arenite, being composed dominantly of sericite and chlorite material, and up to 40% of very fine quartz grains (0.015mm). The sericite-chlorite material is more crystalline than the matrix of the arenite, and shows aggregate polarization in the direction of the slaty cleavage. Detrital muscovite up to 0.1mm (33342) is oriented parallel to the bedding.

Specimen 33316 is without the slaty cleavage and the small sericite flakes show a random orientation. Detrital muscovite comprises about 2%.

Quartzite at Penguin

Between Dial Point and Beecraft Point at Penguin is a series of quartzite beds 100 feet thick interbedded with red argillite. Although surrounded by younger rocks, the structural evidence indicates that they are part of the Burnie Formation.

The quartzites are well-bedded, hard, massive rocks, white in colour with a brown staining due to weathering of the hematite accessories. Specimen 33352 is typical, it contains 79% quartz grains, 19% clay in the form of rounded aggregates and interstitial matrix, 2% chert fragments and accessory tourmaline and rounded hematite.

The quartz grains range in size from 0.2mm to 0.4mm, are equidimensional, tightly packed with strongly sutured borders (Plate 10-a). Some original outlines are still preserved by the presence of secondary enlargements. Chert occurs either as rounded grains up to 0.4mm, or as squashed particles interstitial to the quartz grains. Clay aggregates (0.4mm) also occur as squashed interstitial material, and are difficult to distinguish from clay matrix. The matrix probably does not amount to more than 10%. These are protoquartzites according to the classification of Pettijohn (1957, p.291).

DEPOSITIONAL FEATURES OF THE QUARTZ WACKE-SLATE ASSEMBLAGE

Bedding

The outstanding feature of the Burnie Formation is the rhythmical alternation of a single sedimentation unit of quartz wacke and weakly laminated black mudstone (or slate). The quartz wacke units vary in thickness from a few inches up to four feet. Individual arenite beds can be followed with uniform thickness for up to 300 feet

across the shore platform, giving the impression of great lateral continuity. At some places, for example, Sulphur Creek and Chasm Creek, the rock-type consists of a series of arenite units without mudstone interlayers. At other places the rock-type consists of horizons several hundred feet thick of predominantly black mudstone with abundant small lenses of cross-bedded siltstone.

Festoon cross-bedding

Festoon cross-bedded scoops are very common in the quartz wacke between Round Hill Point and Sulphur Creek. In these deformed rocks, they appear as undulating concentric shear joints (Plate 32-a), and Burns (1964, p.148) has termed them pseudo-boudins. Even when deformed, they can be recognised by the smooth downward-curving parting planes and sharp upward cusps.

The structures are usually confined to the middle of the arenite beds and are thus not sole marks due to scour. They are, however, best seen on the under-side of the beds which have split along the undulating parting planes.

Plate 10-b from the shore platform opposite Sulphur Creek, Siding, and Plate 11-a from a railway cutting 400 yards east of Chasm Creek, show the under-side view of some festoon scoops, which are not modified by deformation. In both examples, the bedding is steeply dipping and slightly overturned. The structures are squat, downward-

lobate and elongate, with a deep, bulbous end and a wider flared end. They vary from 10-20 inches in length, and 5-15 inches in width. Smaller scoops are present within the larger ones, Plate 11-a. The casts display crescentic markings on the lower surface which are approximately parallel to the curvature of the flared end. These are the traces of the cross-laminae infillings on the bottom of the scoop.

The top surface of arenite beds containing these structures consist of closely nestling scoops arranged en échelon, and separated by linguoid ridges which are concave toward the bulbous end. They are thus similar to linguoid ripples.

Precise palaeocurrent determinations have not been attempted using these structures because little is known about their internal structure or their hydrodynamic significance. If they are scour-and-fill structures, then the cross-laminae are interpreted as high angle lee-slope build up, whereas if they are linguoid ripples the cross laminae represents the stoss-slope build up. There is thus a possible error of 180° in using these as palaeocurrent indicators. Furthermore, the closely nestling en échelon nature suggests the water current was "braided", and such features as the elongation of the scoop, the dip direction of cross lamination, and the plane of symmetry would have no simple relationship with

the overall current direction.

Also, the process of unwinding sedimentary lineations in a repeatedly folded sequence involving steeply plunging fold axes, regional overturning, and a flattening component to the strain is difficult (Cummins, 1964), (Ramsay 1961). Nevertheless, an attempt has been made to deduce qualitatively the direction of the currents responsible for these structures by making two assumptions.

The trend of the elongation of the scoop from the bulbous to the flared end is taken as the current direction. This direction is constant in a single exposure and is probably significant. It is also assumed that during structural deformation the major and minor fold axes were initially formed with a rectilinear northeast-southwest trend. This allows a comparison to be made between the direction of current flow and the tectonic lineation, which is readily discernible by the intersection of bedding and cleavage. The sedimentary lineations need not then be unwound by complicated and dubious techniques. This second assumption is most probably sound, despite the structural complexity, because all minor structures are approximately homo-axial due to repeated deformation about a prevailing northeast-southwest axis.

Observations from many localities from Sulphur Creek to Coeee Point indicate that the current directions, although variable, always lie at a high angle to the tectonic lineation. This implies that the currents responsible for the festoon cross-bedding moved approximately along a northwest-southeast line, and therefore were transverse to the northeast-southwest trend of the basin.

Transverse ripples

Transverse ripples occur on the top of some arenite beds at Round Hill Point. These are regular rectilinear corrugations with wavelengths from 2 to 5 inches, and amplitude from 1 to 2 inches, (Plate 11-b). The three examples observed have axes exactly parallel to the first generation tectonic lineation. In the overlying fine-grained siltstone, cross-laminae infillings between crests of the ripples prove a sedimentary origin. These laminations do not simply bend around the corrugations, but fill the depressions with upwardly truncated cross-laminae. The ripples are slightly modified, with cleavage channelled downward into the troughs producing rounded crests and downward cusped troughs. These transverse ripples also point to the presence of transverse traction currents.

Sole-markings of the arenite

The main type of markings on the sole of the arenite beds are flute marks. They are invariably modified by load casting and cleavage development (Plate 12-a), so that the original form is difficult to reconstruct. At Sulphur Creek, Burns (1964, p.28) has described an example of unmodified flutes which are elongate parallel ridges, bulbous at one end and flared at the other. No systematic orientation data of these flutes has been recorded because of the difficulties in recognising the bulbous-to-flared direction and also because of the uncertainties in unwinding palaeocurrent directions through large angles about steeply plunging fold axes in which there has been a flattening component to the deformation. Almost everywhere along the coastline from Howth to Round Hill Point the flute marks are parallel to the tectonic lineation. Assuming that the trend of the current producing the flute is given by the elongation, it is apparent that the currents were directed northeast-southwest along the presumed axis of the trough.

Small-scale scouring of the mudstone at the base of the arenite is common, and is well illustrated in Plate 12-b. Burns (1964, p.28) has described some unusual sole markings from Sulphur Creek.

Graded Bedding

The arenite beds in the rhythmically alternating, thinly bedded, arenite-mudstone lithology, commonly show upward grading from coarse to fine. The base of the arenite bed is invariably sharp and the underlying lutite is scoured. The gradation at the top of the arenite bed into lutite is diffuse in hand-specimen and is commonly accompanied by a fine internal lamination. The gradation generally takes place over a short distance of $\frac{1}{2}$ - 2 inches toward the top of the arenite bed, so that the major part of the arenite is of uniform grain-size. Grading on the microscopic scale is also present, (Plate 12-b).

Internal lamination

The arenite beds generally possess a fine, planar, internal lamination which is due to variation in grain size and matrix content. These internal laminae also show repeated micro-grading and internal load casting (Plate 12-b). Internal lamination in the thicker arenite units is often confined to the top portion, the lower portion being massive.

Small-scale wavy cross-lamination is commonly found in the top few inches of an individual graded arenite unit, and in places is oversteepened and slightly convoluted. This is the type frequently noted in graded graywackes.

Mudstone inclusions

Mudstone inclusions are not everywhere common but are locally abundant in the individual graded arenite beds. They vary in size from 10 inches in length down to a microscopic scale. The larger fragments are generally elongate and lie flat in the bedding plane, but the smaller flakes lack any preferred orientation. These flakes have the peculiar appearance of being sharply angular yet twisted and shredded. The flakes are identical in composition to the interbedded lutite. There is no indication that they are the result of in situ disruption of a once continuous layer, and are interpreted as due to scouring of the newly deposited lutite further up the slope followed by transport and deposition in its present place.

Many of the depositional structures occur in a sequence within the one arenite bed. The thinner arenite beds which average 3 inches in thickness have a sharp basal contact with associated scouring. Above this there may be a rapid gradation into the normal, uniformly grained, featureless rock-type. Following this is an interval of small-scale cross-lamination which may be convoluted. This passes upward into a graded interval with planar lamination, whereby the rock passes into the lutite. The intervals of planar lamination are in places the result of fine, repetitious grading.

The thicker arenite beds up to six feet thick have a sharp basal contact, generally without flutes. The bulk of the bed contains either faint widely spaced planar lamination, or large-scale festoon cross-bedding. The planar or cross-lamination, which is commonly accentuated by concentric shear joints, is a layering 1 to 20 cm in thickness and is due to faint grain size variations. It is not due to repetitious grading. Above this is an interval of planar lamination passing sharply up into the lutite.

The depositional structures indicate that the arenites in the Burnie Formation form two lithogenetic types, a traction deposit characterised by large-scale cross-bedding, and a turbidite deposit characterised by graded bedding and small-scale cross-lamination. Both deposits are ^{microscop}petrographically identical. A detailed analysis has not been attempted, but it appears that the traction deposit is volumetrically more important than the turbidite deposit. This is discussed further in Chapter 6.

|| ?
How can
they be?

INTERBEDDED PILLOW LAVAS

Interbedded pillow lavas occur on the shore platform one mile west of the headland at Sulphur Creek. Similar rocks also occur 15 miles to the west, at a fore-shore outcrop 3 miles west of Somerset.

At Sulphur Creek the lavas are deeply weathered, and the little that is left of the original minerals is

strongly deuterically altered. However, the pillow structure is sufficiently well preserved to prove their volcanic origin. There are *two main* horizons of volcanics, all of which are repeated several times along the shore platform by folding. (Figure 7). They are essentially concordant bodies, varying in thickness from a few feet up to 100 feet, although shearing accompanied by displacement along contacts gives them the appearance of discordant intrusive bodies of variable thickness.

The pillows (Plates 13-a and 13-b) are mostly simple bun-shaped bodies with a flat bottom and a convex top. The extremities of the base are commonly pinched into tails by adjacent pillows, and the larger pillows are draped over the smaller ones. Concentric rows of vesicles are present across the top portion of many of the pillows. The black ribbons between the pillows are composed of a chloritic mudstone similar to the mudstone of the host rock. Despite the fact that the lavas are folded and sheared, it is possible to deduce facings from the pillows, which everywhere agree with the facings of the surrounding sediments obtained from sedimentary depositional structures.

In hand specimen, these lavas are light-green to buff in colour, commonly with small discoid bodies up to 3mm diameter. They are composed of a fine near-isotropic green chlorite, with some sericite and minor

quartz, (33333, 33336). Abundant small granules of epidote are scattered throughout. The discoid bodies are composed of a cloudy mass of finely granular, high-relief barely resolvable material stained by leucoxene. It may be "saussurite" composed of minute calcite, epidote and chlorite. Specimen 33336 is composed of 40% of lath-shaped aggregates of clear sericite with random orientation. These aggregates themselves are up to 0.45mm in length, are twisted and fractured, and possess a good parallel orientation. These may be pseudomorphs of plagioclase with a flow structure. Specimen 33348 is a mottled lava with veins and cellular encrustations of subhedral calcite in a chloritic groundmass.

The rocks west of Somerset are less altered but more deformed than those at Sulphur Creek. They occur in two concordant bodies each about 30 feet thick. These rocks contain flattened and rounded bodies up to 3 feet in diameter which are separated by ribbons of chlorite material. There is, no clear evidence of pillow structure.

Specimens 33313 and 33311 are composed of a chloritic and sericitic matrix with aggregate polarization in the direction of the main cleavage. Scattered throughout are colourless tremolite and fragmented crystals of clinopyroxene up to 1 mm which are strewn out in the foliation. Secondary talc occurs replacing the amphibole. Skeletal ilmenite altering to leucoxene, and shreds of spongy "saussurite" are the accessory minerals.

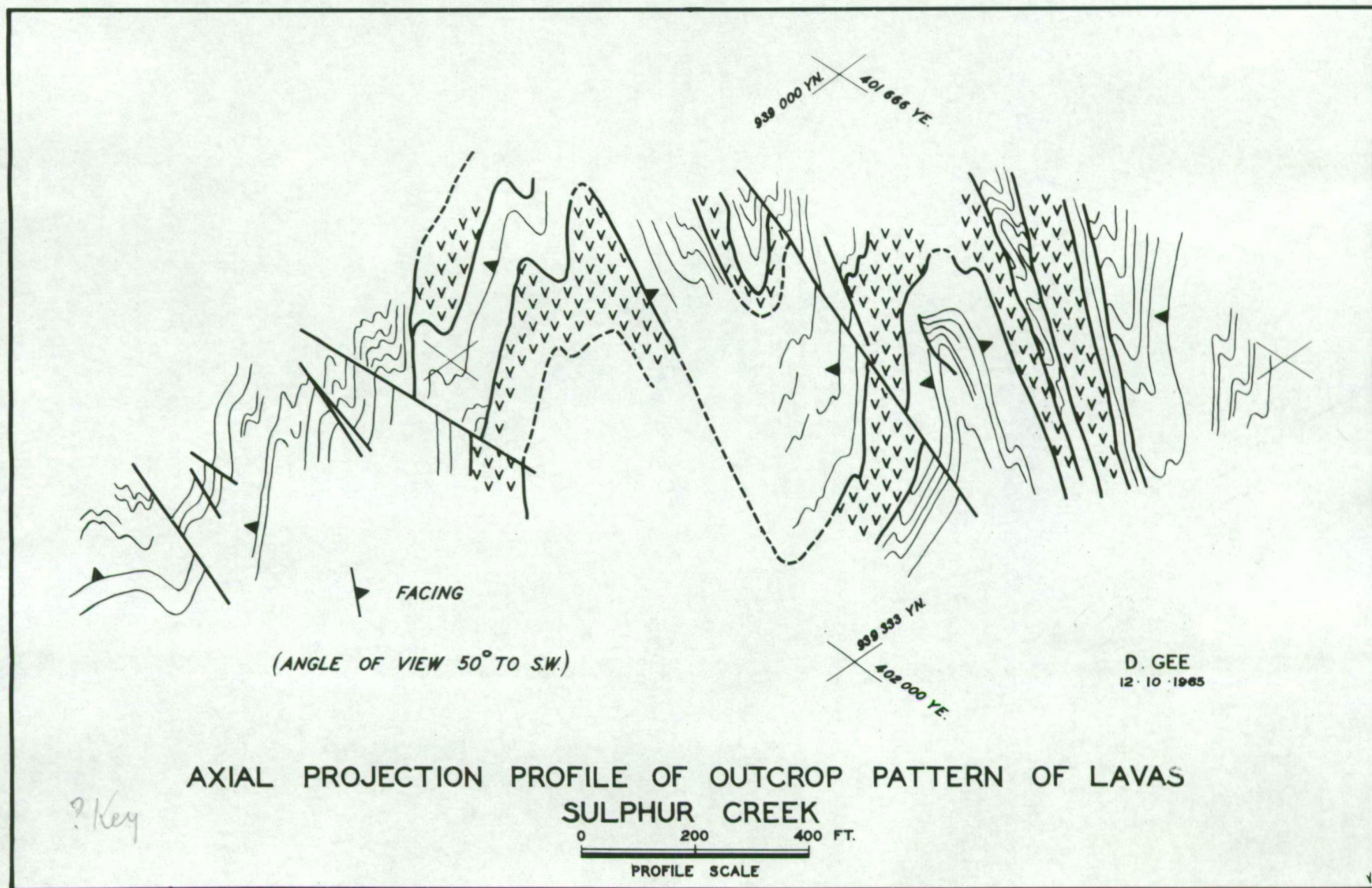


Figure 7.

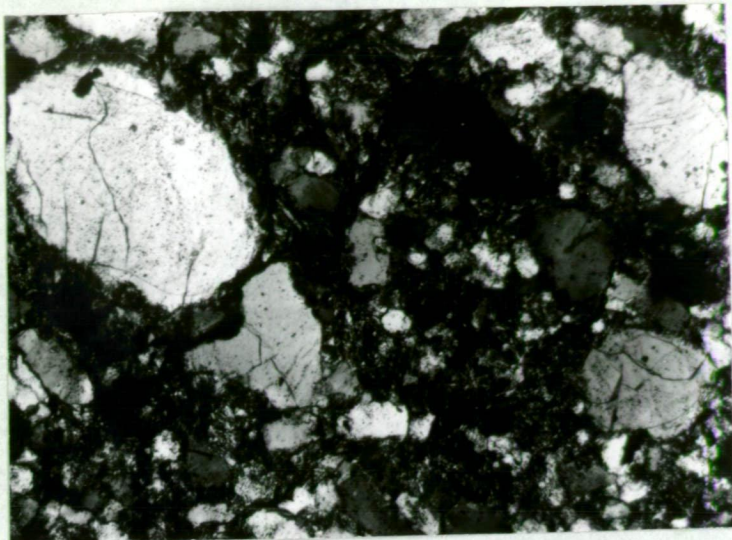


Plate 9a Photomicrograph of corroded overgrowths on initially rounded quartz grains, coarse arenite, Burnie Formation, Sulphur Creek.
Sp. 33340, x 75.

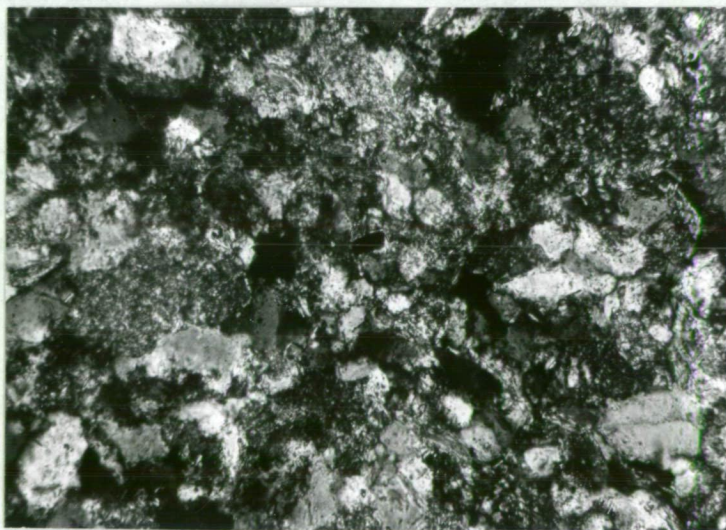


Plate 9b Photomicrograph of typical arenite from the Burnie Formation, showing corroded detrital quartz and squashed chert fragments.
Sp. 33320, x 125.

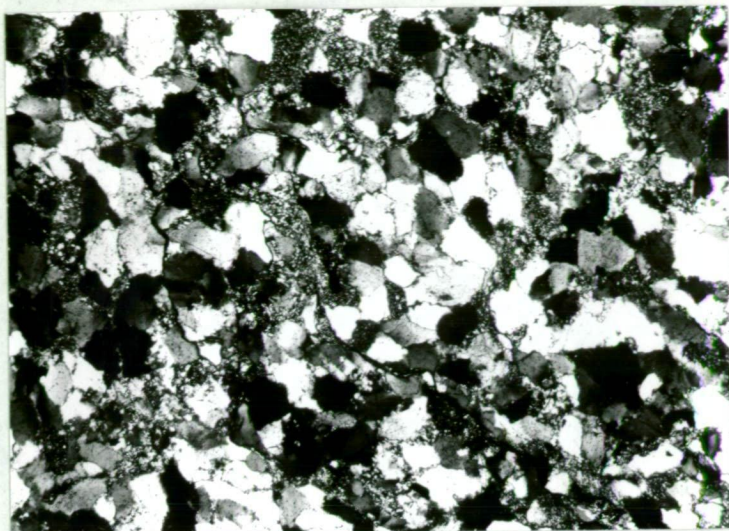


Plate 10a Photomicrograph of quartzite at
Penguin. Sp. 33352, x 25.



Plate 10b Under-side view of festoon cross-
bedding, shore platform, opposite
Sulphur Creek Siding.



Plate 11a Under-side view of festoon cross-bedding, railway cutting, 400 yards east of Chasm Creek.



Plate 11b Rectilinear transverse ripples, Round Hill Point.



Plate 12a Sole marking on an arenite bed,
showing penetration of cleavage
into flames, Chasm Creek.

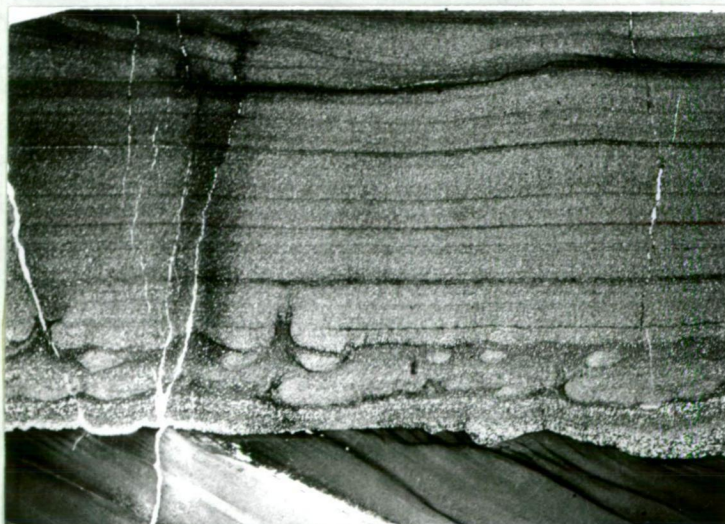


Plate 12b Photomicrograph of a single
arenite bed, showing scours at
base, grading, load casts, inter-
nal planar lamination and cross-
lamination, Round Hill Point.
Sp. 33330, x 2.



Plate 13a Pillow lavas at Sulphur Creek,
top to the right.



Plate 13b Pillow lavas at Sulphur Creek,
top to the foreground.

C H A P T E R 5

SEDIMENTARY DEFORMATION STRUCTURES

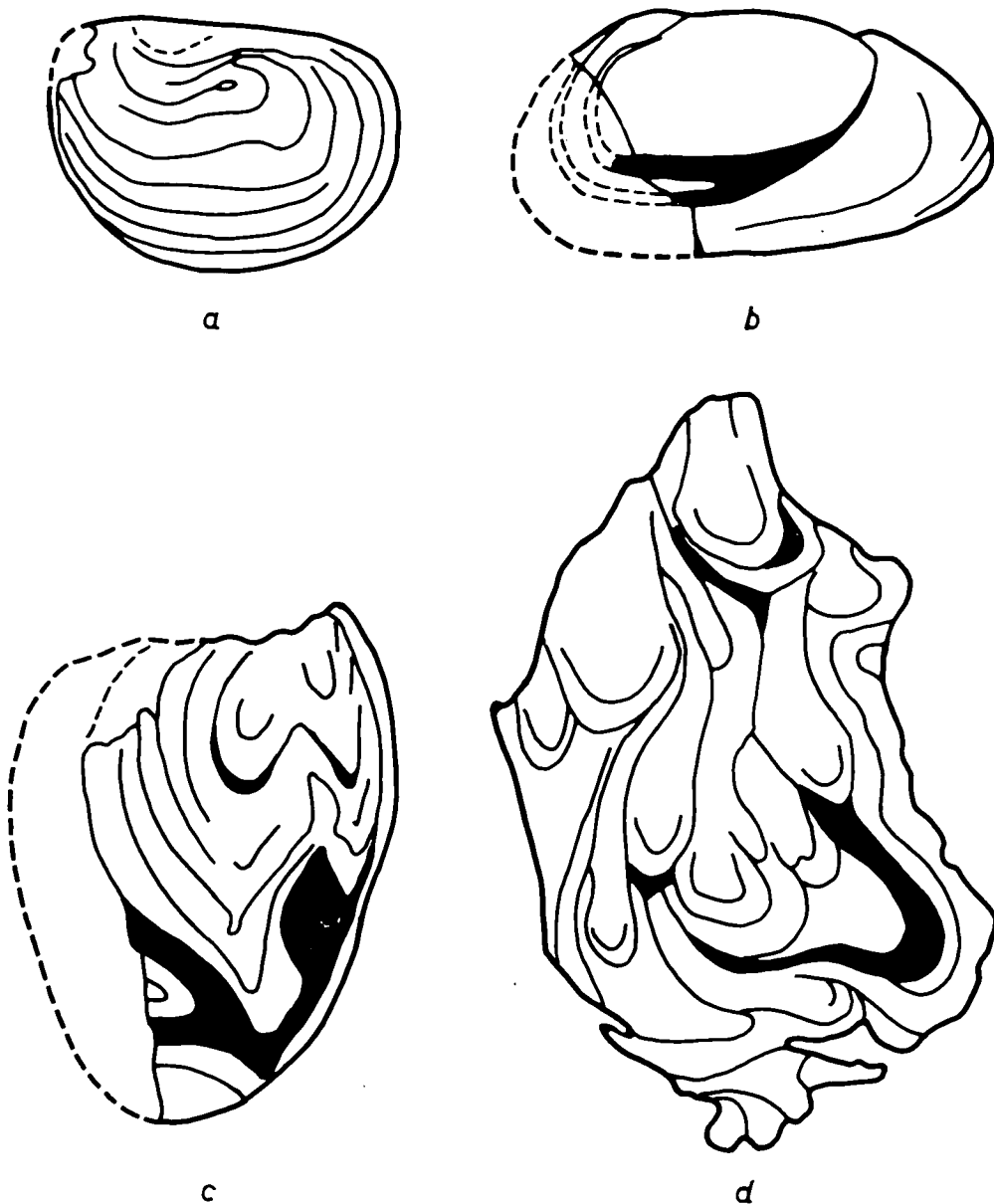
A variety of penecontemporaneous deformational structures is found in the Proterozoic succession in northwest Tasmania. These are described, mechanisms of formation suggested, a genetic classification outlined and criteria suggested to assist in their distinction from tectonic structures.

LOAD CASTS

Load casts are downward lobate protruberances on the base of an arenaceous bed that overlies a lutite bed, formed by the vertical adjustment of the basal material to unequal loading. Either some pre-existing irregularity in the distribution of the overlying arenite, or a variable response of the lutite to equal loading is necessary. Flute marks in the Burnie Formation are invariably accentuated by load casting, (Plate 12-a). These load casts are accompanied by flame structures which are sharp penetrations of mudstone up into the sandstone. Load casts that appear to follow from ripple marks are found in the dolomitic horizon of the Irby Siltstone.

PSEUDO-NODULES

Pseudo-nodules are rounded bodies of sandstone embedded in a mudstone layer. They are the result of



SECTIONS OF PSEUDO-NODULES

a, b, c, simple pseudo-nodules from unnamed siltstones at Jacobs Boat Harbour; d, complex pseudo-nodule from Cowrie Siltstone, west side of Rocky Cape. Natural Scale. Lines and solid shading indicate thin claystone laminae.

Figure 8

advanced load casting, during which a sand layer is broken into detached bodies.

Excellent examples of pseudo-nodules are found in the unassigned sandstone and siltstone formation 300 yards east of the headland at Jacobs Boat Harbour. The bedding is irregular and the sandstone units vary from 2 inches to 2 feet in thickness. The base of the thinner sandstone layers have well developed load casts sagging into the underlying siltstone and are accompanied by flames. All stages are observed in the disruption of a wedging sandstone unit, and Plate 14-a illustrates the end product. Detached pseudo-nodules occur where the bed thickness decreases below six inches. In cross-section they are roughly ellipsoidal and in the third dimension are either cigar-shaped or equidimensional. Internally, (Figures 8-a, 8-b and 8-c) they have a simple concentric structure of sandy laminae 2 - 5mm thick, separated by paper-thin films of mudstone. These laminae form a continuous wrapping around the bottom half where they are thicker, but are absent at the top.

Figure 8-d is a direct tracing of a complex pseudo-nodule from the Cowrie Siltstone on the west side of Rocky Cape. This pseudo-nodule consists of a group of closely nestling, downward-sagging, droplet-shaped lobes of laminated sandstone. The lamination is once again due to paper-thin layers of mudstone. The sandstone

laminae are thicker at the bottom of each lobe and are attenuated further up the tail. The contortions at the bottom are formed first and are bent over and squashed by the succeeding droplets. This complex structure is considered to result from sagging of several thin layers of sandstone separated by thin layers of mud, together contained within a thicker bed of mud. As lobes are formed, most of the interlayered mud is squeezed out. All stages in the break-up of the initial laminae are present, and when complete, the horizon is marked by a row of pseudo-nodules.

The internal structure of the simple and complex pseudo-nodules is a clear indication that vertical movement due to gravity alone was responsible. Kelling and Walton (1957) adequately explained the origin of load casts by a hydrostatic adjustment of sand of density about 2.1, overlying silt of density about 1.4. Small irregularities in thickness of the sand load resulted in differential hydrostatic pressures along the sand-silt interface which are sufficient to cause flowage in the unconsolidated sediment. Keunen (1958) experimentally produced almost identical structures, by allowing sand to sink down into thixotropic mud. The upward piercement of the finer grained material and the sagging of the coarser material results in the curling-up of the sandstone laminae. It is important

to note that complex contortion and extreme attenuation of the bedding lamination can result without any lateral down-slope movement.

DIAPYRIC CONTORTION

A peculiar type of contorted stratification is present in the Jacob Quartzite on the headland at Jacobs Boat Harbour. The rock-type, as previously described is a mature quartz sandstone, composed of well-rounded and well-sorted grains of medium sand size. The rock contains no matrix. It has been suggested that this quartzite was deposited in shallow, current-swept waters in, or just beneath, the littoral zone.

The contorted stratification is confined to a discrete zone 30 feet thick, which is parallel to the bedding. The zone can be traced for 800 yards down the limb of the large syncline just west of Boat Harbour, and a contorted zone $1\frac{1}{2}$ miles further to the west, on the opposite limb of the syncline, is probably the same zone.

In profile the laminae are contorted and twisted, although the bedding units, which average 12 inches in thickness, are still discernible. The gross aspect of bedding is therefore the same as the normal bedding, (Plate 14-b). The plications take the form of rounded synclinal depressions, and sharp cusped anticlines.

The axial surfaces are generally perpendicular to the gross aspect of bedding, and there is no overturning. The amplitude of the plications rarely exceeds 3 inches.

When viewed from above, the bedding surfaces are patterned with saucer-shaped and amoeboid depressions, separated by ropy fold crests. The fold axes generally have a random orientation within the bedding plane but commonly are tipped up at an angle to the bedding. The symmetry is therefore broadly polar, indicating upward diapiric movement. These rocks are normally rich in cross-bedding, but it is rare in the contorted zone.

The upper contact of the contorted zone is quite sharp, and follows the same stratigraphic interval over a distance of at least 200 feet. The immediately overlying bed is cross-bedded. At the lower contact the folds die out downward within the bottom 12 inches onto a thin bed of siltstone. There is no evidence of any plane of detachment at the base.

In suggesting the origin, the following points are important:-

- (1) the mature, well sorted rock-type;
- (2) the presumed littoral or sub-littoral environment;
- (3) the confinement of the layer parallel to bedding;
- (4) the preservation of gross aspect of bedding;

- (5) the rounded synclines and cusped anticlines;
- (6) the polar symmetry;
- (7) the absence of cross-bedding in an otherwise cross-bedded sequence;

The deformation is considered to be due to the upward escape of entrapped interstitial fluids. No large-scale slumping or dislocation is involved. The sands were probably laid down in quiet shallow water where waves and currents were not sufficiently strong to form cross-bedding or ripple mark. The environment may have been a tidal lagoon. This structure does not appear to be the result of simple load deformation due to density differences because there are no siltstone layers within the deformed zone.

There are two possible origins. Stewart (1956) described similar contorted stratification in a modern coastal lagoon, thought to be due to air heave. The sediments described by Stewart were deposited in a protected tidal lagoon with the flood tide, producing a layer of cavernous sand. On the ebb tide, air is sucked into the voids, and on the next flood tide this air is trapped in a narrow zone between the water above, and the rising water table within the sand. Stewart (1956) was able to show experimentally that this upward movement of entrapped air was sufficient to cause deformation.

Because of the large thickness of sediment involved, this process must have operated continuously, with a prolonged and delicate balance being maintained between the rate of sedimentation and the energy of environment. No counterpart of the miniature unconformities noted by Stewart is found in the rocks at Jacobs Boat Harbour.

The second, and more likely possibility is that deformation occurred while the sediment was in a state of temporary liquefaction caused by a collapse of unstable cavernous or honeycomb textures in the newly deposited, waterlogged beds, as outlined by Terzaghi (1957) and Williams (1960). Such a collapse may be due to the normal process of compaction under its own load. Such a collapse would result in a momentary but substantial lowering of shear resistance, expulsion of water, and the formation of a more closely packed and stable texture. In this state the granular aggregate would be dilatant so that the viscosity would markedly increase and deformation would be arrested.

BALL-AND-PILLOW STRUCTURE

In the Detention Sub-group, exposed in the railway cuttings through the Sisters Hills, is an unusual type of penecontemporaneous intraformational disturbance resembling structures that have been called ball-and-pillow structure. (Potter and Pettijohn, 1963).

The rock type is a regularly bedded, flaggy quartzite with orthoquartzite affinities, interbedded with minor thin siltstone beds. Ripple mark is present.

The disturbances are restricted to zones varying from 2 feet to 8 feet thick, and which contain more than one sedimentation unit. The zones are parallel to the bedding and are separated by up to 50 feet of undisturbed beds. In one railway cutting 100 feet long, five of these zones were counted, each about 5 feet thick.

Internally, the disturbed zones consist of tightly packed, detached, isoclinal fold cores which tend to be pillow-shaped. The contortions are almost entirely downward lobate synclines and corresponding anticlines are rare. Where present, the anticlines are very acute. (Plate 15-a). In the thinner deformed zones the upturned limbs of the synclines are truncated by the overlying undisturbed bed. The anticline in Plate 15-a has a variable plunge. In a distance of 18 inches it steepens in plunge from 0° to 40° , and then in another 6 inches it vanishes into a broad upward undulation which then forms a planar surface.

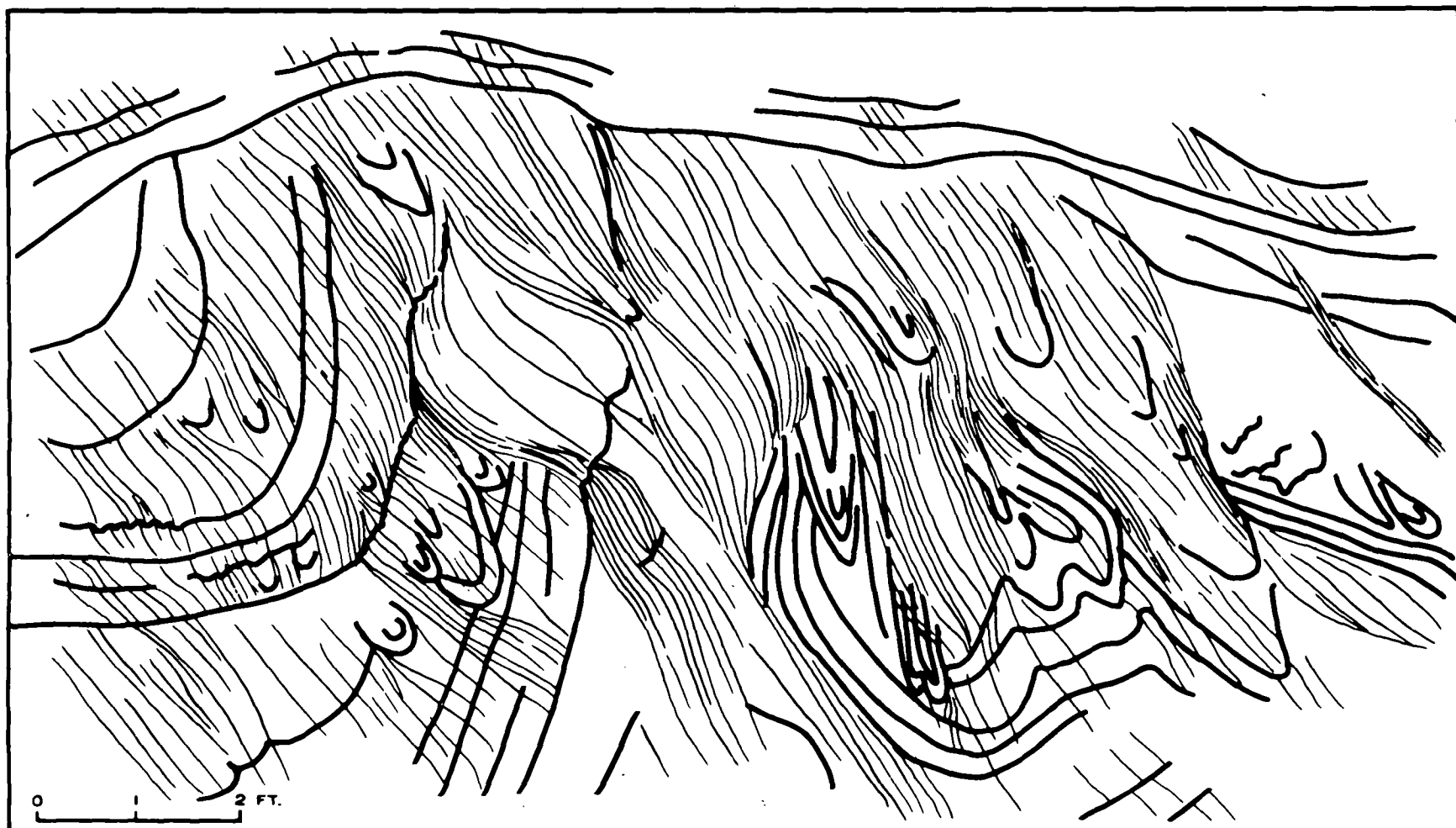
In most cases the axial planes are inclined at a high angle to the bedding, so that after removal of tectonic tilt the axial planes are approximately vertical. Within the one disturbed zone, the fold axes

have approximately the same azimuth but with a variable plunge. Between outcrops there is considerable variation in azimuths.

The best exposure of these structures is in the rail cutting at the $138\frac{3}{4}$ mile peg. Here the disturbed bed is 9 feet thick. Figure 9 is a sketch drawn in the field from a series of photographs, showing the style of folding, the mutual relations of the folds and the relationship of the tectonic cleavage.

The larger, more open synclines are concentric in style and affect several sedimentation units. The laminae within these sedimentation units are deformed into smaller isoclinal synclines generally with smooth outlines. The laminae are thicker around the crests of these folds, and thin rapidly up the limbs. Four of these smaller synclines in the middle-left of Figure 9 are closely nestling structures with the upper ones intruding down into the lower ones. The intruded fold cores are not distorted and there is no brecciation or slickensiding.

The bedding laminae themselves, which are due to slight grain-size variations and small amounts of heavy minerals, are usually well preserved. However, there are isolated patches of featureless granular sandstone in which the laminae appear to have been obliterated. These patches are generally associated with



BALL-AND-PILLOW STRUCTURE, SHOWING DISPOSITION
OF TECTONIC CLEAVAGE. DETENTION SUB-GROUP QUARTZITE,
SISTERS HILLS RAILWAY CUTTING 138 $\frac{3}{4}$ MILE PEG

pockets of smaller scale and more irregular contortion where the lamination is not quite obliterated.

These structures bear a special relationship to the tectonic cleavage, and could possibly be interpreted as a peculiar type of folding induced by the regional folding. The cleavage is deflected around the rounded synclines and channelled up into the cusped anticlines. The net result in some beds is a strongly schistose rock with an undulating cleavage which encloses unsheared and detached synclinal fold cores. In this locality the cleavage is at a high angle to the bedding, thus the cleavage plane lies close to the axial planes of the lobes. In this respect it could be argued that this structure is the lobate bedding of Hills (1963, p.300, Figure X-16), which he thought to be due to the upward escape of interstitial water along cleavage planes during folding. This is supported by the fact that all these disturbed zones in the railway cuttings occur in the crest of a major anticline where entrapped connate waters may be expected to accumulate, after the initial folding, and before cleavage development in the crest. However, it is suggested that this relationship between cleavage and lobes is fortuitous, and that the lobes were present before cleavage development. There are cases where the lobes are recumbent and the tectonic cleavage cuts obliquely across the axial planes. One example is illustrated in Figure 10.

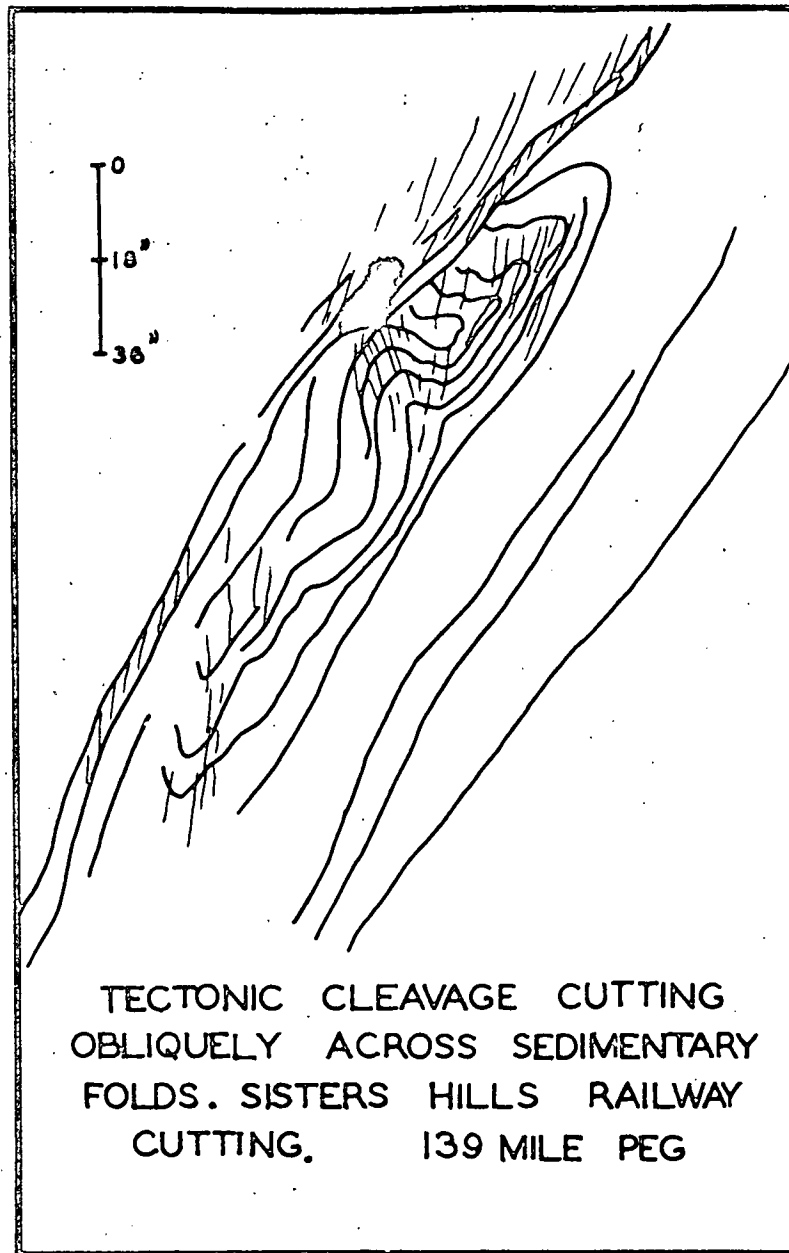


Figure 10

The form and orientation of the structures show that mainly vertical movement is involved and any lateral movement is negligible. The normal mechanics of load casting are not applicable here because of the lithological uniformity within the disturbed zones, and because of the large size of the structures. The upper contact of the disturbed layer is slightly irregular and sharp, and truncates the upturned structures. This establishes that the structures were formed before deposition of the overlying material, in which case they result from some type of underwater disturbance of newly deposited sand.

The origin is probably similar to that suggested for the diapiric contortion at Jacobs Boat Harbour although the structures differ in some respects. At Jacobs Boat Harbour the gross aspect of bedding within the disturbed zone is still recognisable, the individual folds are smaller in size and the anticlines are still preserved.

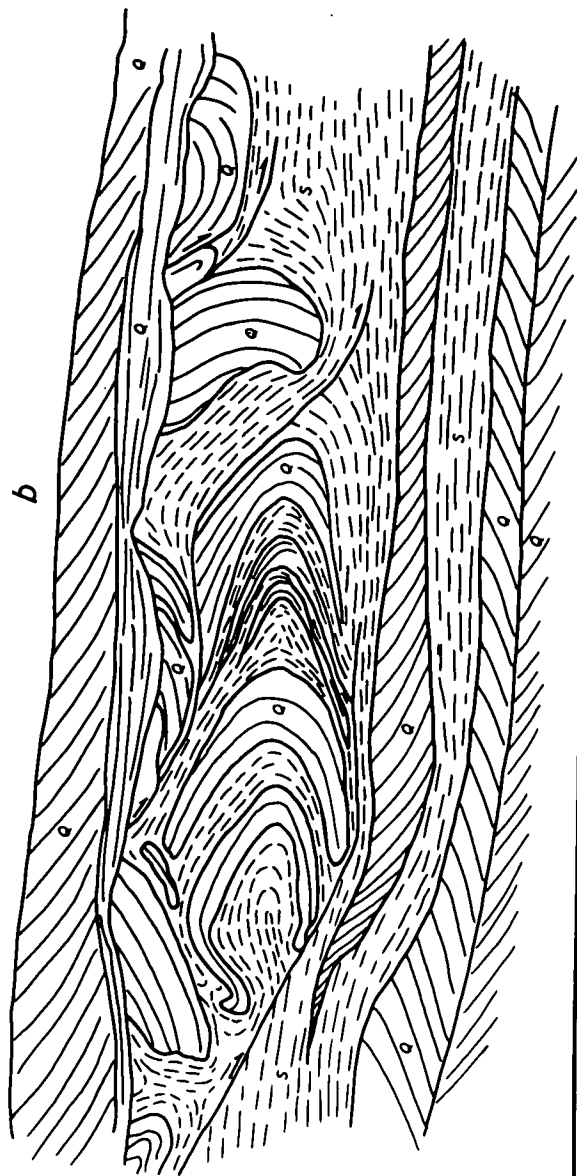
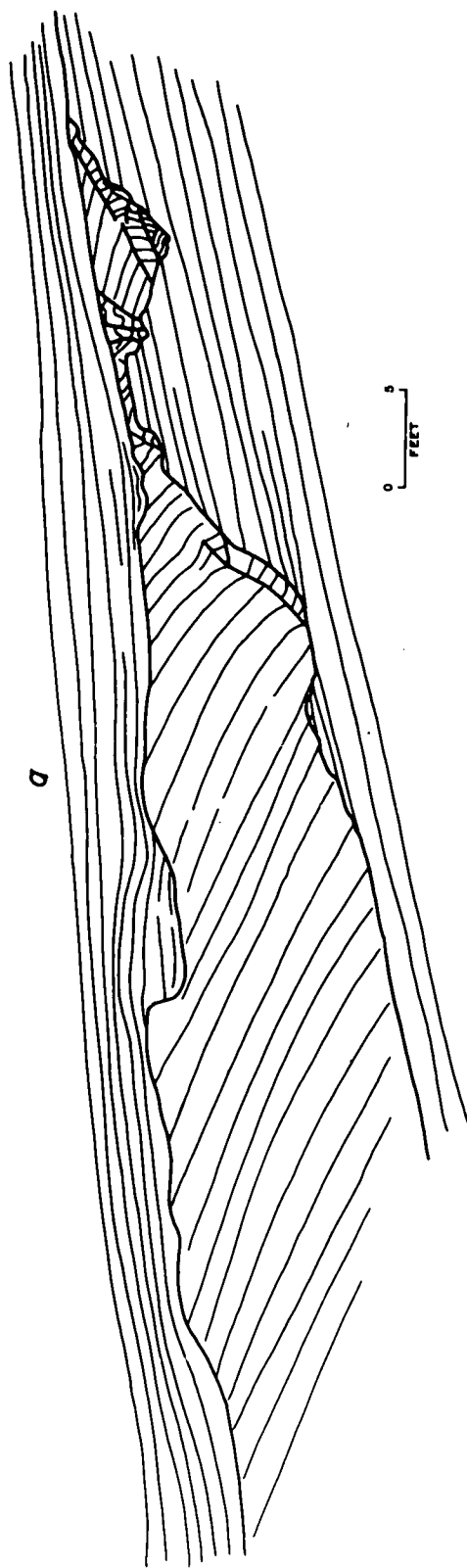
The morphology and symmetry of the ball-and-pillow structures indicate that they must have also formed by the upward movement of interstitial fluids in newly deposited sediment during compaction. Selley et.al. (1963) have described similar structures and suggested that they are the expression of movement in ancient quicksands.

If the structures at Jacobs Boat Harbour and the Sisters Hills railway cuttings have a common origin then there must have been a difference in behaviour of the sediments. The reason for this is uncertain. The deformations in the Sisters Hills were initially larger and proceeded further, suggesting a lower viscosity for longer periods. One possible explanation is that the rocks in the Sisters Hills railway cuttings contain up to 10% clay fraction. Boswell (1961, p.74) notes that the addition of a small quantity of clay to a dilatant sand makes it thixotropic.

BLOCK SLIDE

A slide is a structure formed by mass movement of a semi-consolidated sediment along discrete shear planes, accompanied by relatively minor internal plastic flow and variable amounts of fracturing.

About 300 feet from the top of the Cowrie Siltstone, on the west side of Rocky Cape, is an excellent example of a block slide. Figure 11-a is a field sketch of the slide. The block is 12 feet thick, and is exposed over a length of 80 feet, but the total length may be much greater. Bedding within the block is sharply discordant to the normal attitude of bedding, showing a bodily rotation as well as a lateral displacement of the block. The sole plane, under the main part of the



- a Block-slide in laminated shale
Cowie Siltstone, western side
of Rocky Cape.
- b Slump-sheet in silty lens in
the cross-bedded Jacob
Quartzite $\frac{1}{2}$ miles west of
Jacobs Boat Harbour.
Quartzite-Q, Siltstone-S

Figure 11.

block, is smooth and follows the bedding underneath. In detail (Plate 15-b), the basal plane is made up of a nest of gently undulating surfaces which define a shear zone two inches in width.

Toward the heel, the basal slide surface turns up in a double shear circle. Here, the displaced block clearly truncates the underlying beds, dragging them slightly downward. Still further back, the movement occurs on irregular and complex surfaces that branch and transgress back up through the underlying beds. In this zone the bedding is fractured and crumpled. The deformed block then disappears due to truncation by the overlying beds. Much of the larger scale irregularity of the bottom contact in Figure 9-a, is due to an oblique view of the section. The sketch is drawn looking north and the direction of movement (deduced below) is to the northwest.

The upper surface is a smooth or slightly irregular plane upon which the overlying undisturbed beds were deposited with angular discordance. The overlying beds are of non-laminated cross-bedded siltstone and fine sandstone, distinct from the laminated mudstone in the displaced block.

Bedding in the displaced block is crumpled only in the fracture zone. The block has been displaced

by sliding and by a rotation of about 15° . The axis of this rotation, given ^{by} the line of intersection of bedding within the block and beneath the block, is 030° . This is also the axis of curvature of the shear circles. The direction of movement is approximately perpendicular to the axis of rotation, in a northwesterly direction.

This structure has features directly analogous with subaerial land slips, and is therefore probably the expression of downslope submarine block sliding under the influence of gravity. The slope was toward the northwest. Movement was probably caused by removal of support at some distance downslope, or by a slight tilting which accentuated the depositional slope.

SLUMP SHEETS

The terms "slumping", "slump structure" and "slump sheet" have been widely and loosely used in the geological literature. Slump sheets were first considered in detail by Jones (1937) who used the term for those structures resulting from downslope gravitational movement involving crumpling with lateral displacement of one or more sedimentation units, at the sediment-water interface. This process was called slumping. Slumping has on occasions in the past been used as a "bag-name" for all soft-sediment contortion. Slumping should clearly be distinguished from processes which

produce load structure, ball-and-pillow structure and convolute lamination.

Slumping is one aspect of a wide range of down-slope subaqueous deformational processes which depend on little known variables such as the degree of consolidation, the cohesion of the sediments (which is controlled in part by the grain size and mineral composition), water content, and the angle of deposition of the sediment. Dott (1963) recognised a wide spectrum of structures, ranging from block slides formed by movement along a discrete shear plane with minor plastic flow; slump sheets with recumbent folds and variable disruption of bedding; mudflows or fluxoturbidites in which internal stratification is almost completely destroyed; and finally turbidity currents.

Examples of block slides, slump-sheets, fluxo-turbidites and turbidity current deposits are present in the Proterozoic succession in northwest Tasmania.

Slump structures are fairly common in the Burnie Formation, and also occur in the interbedded lutite, in the orthoquartzite formations of the Rocky Cape Group. Figure 11-b is an example of a slump sheet in an intercalated siltstone bed in the Jacob Orthoquartzite. It is located three quarters of a mile west of Jacobs Boat Harbour. The slump sheet is confined

to a bed four feet thick of siltstone with interbedded sandstone lenses. The slump sheet consists of a series of small recumbent fold noses, and low-angle slide planes. Movement on these slide planes is the same direction as that indicated by the facing of the recumbent fold-noses.

The top contact in this example is a plane of erosion. The immediately overlying bed is an irregular sheet of sandstone, which is in turn overlain by a typical bed of cross-bedded sandstone. Deformation therefore occurred prior to the deposition of the overlying beds.

The fold axes all lie parallel to the bedding and plunge southeast, and when the bedding tilt is corrected, movement toward the southwest is indicated. Seventeen measurements of cross-bedding were also recorded at this locality (Station 13 on Figure 4), and these have a strong maximum in the southwest quadrant. This does not imply deformation by current drag, but probably means that slumping was directed down the local depositional slope.

A different type of slump structure is found in the Burnie Formation. Examples occur at West Park Point (Plate 16-a), at Round Hill Point (Plate 16-b) and at Freezers Point at Somerset. The slump sheets vary between six inches and 18 inches in thickness and are laterally persistent for distances of up to at least

100 yards. They affect zones of mudstone which contain abundant layers and cross-bedded lenses of sandstone up to 4 inches thick, and thus involve more than one sedimentation unit. Internally, the slump sheets contain abundant tight recumbent, and strongly disharmonic folds. The fold axes all lie parallel to the bedding plane but otherwise have no preferred alignment. The symmetry of the structures indicates lateral movement rather than vertical movement. All show top-truncation and also truncation at the bottom by a sharply defined plane of parting, (Plate 16-b). This bottom-truncation is considered to be dislocation during the later tectonic folding.

SEDIMENTARY SCHUPPEN STRUCTURE

An unusual type of imbricate slumping is found in one horizon of cross-bedded sandstone lenses within the Cowrie Siltstone at Cowrie Point. Such horizons of cross-bedded lenses are common in the Cowrie Siltstone and have been described in Chapter 3.

The lenses are piled one against the other in an imbricate manner, (Plate 17-a), resembling a schuppen structure. Small sigmoidal thrust faults occur between individual lenses, and these dip in the opposite direction to the inferred movement direction. By analogy with schuppen structure the direction of movement in Plate 17-a is from right to left. Movement on the faults

is taken up mostly by tailing out of the small thrusts into ^{the} bedding plane, although there is some puckering of the underlying siltstone. The laminae within the cross-bedded lenses are also sigmoidal and deformed in a manner consistent with the idea of dragging by an over-riding movement.

The lenses are elongate, spindle-shaped bodies, oval in cross-section with maximum and minimum diameters of 18 inches and 9 inches respectively. This is larger than the average cross-bedded lens in the Cowrie Siltstone. A small area of the lower surface has been stripped of sandstone lenses by erosion, revealing a somewhat regularly puckered surface of closely nestling, en échelon spindle-shaped basins up to 4 feet long, separated by cusped and sinuous "domes".

The exact movement picture is difficult to reconstruct as it is not clear whether these lenses were originally separate bodies, or formed the one sandstone bed with a rippled upper surface. The latter case requires less lateral movement, but in both cases the displacement along the bedding is cumulative. Thus, succeeding lenses in a uniformly spaced train of lenses have to slide progressively greater distances in order to form a closely nestling schuppen structure. The amount of relative lateral shortening is probably more than 50% at the present length of exposure, but the actual

amount of movement may only be of the order of 10 feet. Movement of sediment was due to down-slope sliding, either under the force of its own weight, or by drag exerted by a vigorous water current. The fact that the lateral movement must have been cumulative suggests that the movement occurred at the sediment-water interface.

RECUMBENT CROSS-BEDDING

In the Jacob Quartzite rare examples of recumbent cross-bedding are seen, and one example is illustrated in Plate 5-b. This takes the form of a smooth bending of the upper part of the cross-bedded unit in the down current direction. This phenomenon is not rare and Potter and Pettijohn (1963, p.80) refer to many examples. Recumbent cross-bedding has been experimentally investigated by McKee, et. al., (1962) who consider it due to the dragging exerted on saturated cross-stratified sand by a sudden rush of water. McKee and others (1962) also noted that these structures occur mainly in fluvial deposits, in which case the sudden rush of water may be interpreted as floods.

CONVOLUTE LAMINATION

The term "convolute lamination" is used following broadly the original terminologies of Ka nen (1952, 1953b) and Ten Haaf (1956) for sets of wavy contorted lamination having the following properties:-

- (a) sharp anticlines and rounded synclines,
- (b) no external irregularity of the convoluted layer,
- (c) an upward increase in intensity of the contortions which then may die out against the top contact or be truncated,
- (d) continuous laminae around the convolutions,
- (e) absence of contemporaneous faulting,
- (f) possible association of small scours and ripples.

This accords with widespread current usage, for example, Holland (1959, p.233), Sanders (1960, p.418), Douma (1962), Dott and Howard (1962, p.114, p.120), Potter and Pettijohn (1963, p.152), Dzulynski and Smith (1963, p.618) and Davies (1965, p.306). It is distinct from other types of contorted stratification such as slump sheets, load casting, ball-and-pillow structure and pseudo-nodules. Even within this narrow sense, there appear to be two types of convolute lamination in the Proterozoic of northwest Tasmania. Each type is different in morphology and in the association with other structures, but both types probably have a similar origin.

One type of convolute lamination, illustrated in Plate 17-b, occurs in the Cowrie Siltstone, one mile west of Rocky Cape, in a non-graded interval involving

several sedimentation units. The deformed layer contains folded laminae which die out rapidly at the top and bottom, so that the folds are flattened against the confining walls, giving the appearance of box folds. There is no truncation, and it is clearly a closed-cast deformation. The rock type involved is a medium-grained siltstone which is generally finely laminated. In patches, it loses its fine lamination and has a structureless sugary appearance.

This phenomenon is best explained by the mechanism of Williams (1960), involving liquefaction of confined, waterlogged, unconsolidated layers. Liquefaction, as outlined by Terzaghi (1957) is the spontaneous collapse of newly deposited silt with metastable cavernous or honeycomb textures. The collapse results in the expulsion of water and a substantial lowering of the shear resistance. Williams (1960) argues that this expulsion of water from one part of the bed to another part would be sufficient to cause intra-stratal flow of the fluidized sediment. The complexity of the resulting deformation is dependent more upon the relationship of the flow pattern and the sedimentary lamination, than upon any large-scale lateral movement of the overlying beds.

The other type of convolute lamination occurs in the Burnie Formation, within the graded arenite units which are considered to be turbidity current deposits. This takes the form of gentle undulations and occasional overturning of the internal fine cross-lamination which generally occurs toward the top of the graded arenite unit. This convolute lamination has often been described and illustrated in the cross-laminated interval of the graded arenite unit in turbidites, (Bouma, 1962, Dzulynski and Smith 1963, Holland 1959, Sanders 1960, Kojonen 1953), and some of these writers suggest an origin by drag due to hydraulic forces of contemporaneous current action. However, no theory has been advanced that adequately explains either the behaviour of the sediment during deformation (Williams 1960), or the presence of the convoluted layer within the turbidite unit that was presumably deposited very quickly (Holland 1961).

It is possible that some of the examples cited by the above writers may be due to the mechanism suggested by Williams (1960) and later extended to include water-covered, unconfined beds, (1961). Thus where convolute lamination appears to be closely related to structure of contemporaneous current action, open cast liquefaction may have occurred. The close association of convolute lamination and cross lamination may be a fortuitous

relationship, and determined by grain size. Fine-grained sand and coarse silt is the size-grade range most susceptible to liquefaction and this is often the grain size near the top of the graded arenite unit where the individual detrital grains start moving in traction.

CLASSIFICATION

The structures described above fall into two natural classes (Table 3). Firstly there are syn-depositional structures formed by an adjustment of the unconsolidated sediment to the hydraulic flow conditions at the sediment-water interface. This class includes recumbent cross-bedding and also syndepositional convolute lamination (streaked-out ripples) which may not be present in this area.

Secondly, there are post-depositional structures formed by some disturbance within the newly deposited sediment. This can either be closed cast if the disturbed zone is confined above by undisturbed sediment at the time of deformation, or open cast if the disturbance reaches the sediment-water interface. The post-depositional structures can be divided according to the types of movement involved.

- (a) Structures due to lateral mass movement under the influence of a resolved component of gravity acting down an incline. Dott (1963) has classified these structures according to whether the behaviour of the sediment is

elastic, plastic or viscous, thus allowing a continuous range of structures from block slides, slump sheets, fluxo-turbidites, (and turbidites).

- (b) Structures due to small movements in either the horizontal or vertical direction, or various combinations of both directions. Such structures may result from density inversions of layered sediment, upward escape of entrapped connate fluids, thixotropic breakdown or liquefaction. The movements involved are often nondirectional or mutually cancelling so that there is no bulk deformation on the outcrop scale. Thus small horizontal movements may be set up to compensate for vertical displacements during pseudo-nodule formation, producing toroidal or more complex cells.

These movement pictures are different in process, but are not mutually exclusive. For example, some of the pseudo-nodules in the Cowrie Siltstone are drawn out in the bedding, indicating some lateral movement as well as the vertical movement. The precise movement picture is deduced from the symmetry, style and orientation.

CLASSIFICATION OF SEDIMENTARY DEFORMATION STRUCTURES

SYNDEPOSITIONAL		POSTDEPOSITIONAL		
	Movement involving downslope mass-movement	Movement small, often mutually cancelling in different directions		
	block slide slump sheet fluxo-turbidite	vertical	toroidal	horizontal
		load cast	pseudonodule →	diapiric contortion → ball-and-pillow-structure → convolute lamination → (truncated or complete)
syndepositional convolute lamination recumbent cross-bedding				

Table 3.

THE DISTINCTION BETWEEN SEDIMENTARY AND TECTONIC STRUCTURES

It is important to distinguish correctly between penecontemporaneous sedimentary structures and tectonic structures, otherwise a completely wrong interpretation of the structural geometry and hence the tectonic history of an area may be made. In the rocks under discussion, this may be made with some confidence since the sedimentary phase can be distinguished from the tectonic phase. Many textbooks (eg. Twenhofel 1932, p.744; Lahee 1941, p.189; Nevin 1949, p.201; Gougel 1962, p.102; Pettijohn 1957, p.157 and deSitter 1956, p.201) list some criteria which help to make this distinction, but they are not universally applicable. Most have limitations due to the fact that style of folding is not a conclusive criterion since both unconsolidated and lithified sediments can be involved in tectonic deformation. This introduces the difficult problem of the definition of the words "sedimentary" and "tectonic".

Tectonic deformation, often taken to imply deformation of lithified rock, is difficult to define because recent work has shown large-scale subaqueous gravity movements which have transported large masses of unconsolidated sediment over great distances. In complex geosynclines where sedimentation is proceeding concurrently with tectonic deformation it is inevitable

that newly deposited sediments must receive some deformation. It is therefore not valid to say that sedimentary structures are those formed in the sedimentary phase of geosyncline development, and tectonic structures are those formed in the deformation phase. From a study of the relevant literature it is not difficult to present a gradation in size from small gravity slumps up to structures of nappe dimensions. It may therefore be argued that all such structures should be labelled tectonic. However, this is undesirable since structures such as graded-bedding, load casts or convolute folds would then be termed tectonic structures. Some examples of contorted stratification described in this chapter are believed to have formed in an extremely stable tectonic environment and need no initial slope. These structures are related to the rheotropic properties of the newly deposited sediment and not to any short- or long-period stress differences in the sedimentary pile. Structures such as load casts, pseudo-nodules, ball-and-pillow structure, diapiric contortions, convolute lamination and recumbent cross-bedding are clear-cut in concept from tectonic structures. They are small-scale and occur in unconsolidated sediment at, or very near, the sediment-water interface. They may form irrespective of whether the sedimentary basin undergoes deformation.

Subaqueous gravitational slumping is commonly attributed to one or several of the following factors:-

- (a) triggering of an unstable mass by sudden shock due to earthquakes, volcanic loading or impact of storm waves;
- (b) oversteepening of the depositional interface, by prograding of a delta front, or removal of frontal support in submarine canyons;
- (c) a spontaneous breakdown of the sediments which may be by thixotropy as outlined by Boswell (1961), or liquefaction (Terzaghi 1957);
- (d) accentuation of the depositional slope by regional tilting or vertical movements.

It is probable that factors (a), (b) and (c) produce small-scale slumps of the type described by Jones (1937) and described previously in this chapter. Although millions of tons may be involved in the slump sheet, the deformation is superficial and does not constitute bulk deformation of the sedimentary pile. These structures are intimately connected with the deposition of turbidite and deltaic sediments, and slumping is probably a very important agent of sedimentation in some geosynclines, (Heezen, and Drake 1963)

Much larger and more complex structures can develop by process (d) where the sedimentary pile is actively undergoing tectonic deformation. These are the

early tectonic (syntaphral) structures. Thus, Waterhouse and Bradley (1957) described a series of deep-seated recumbent folds in the Mungaroa Limestone of New Zealand, thought to represent down-slope "lurches" in the unconsolidated limestone following warping and tilting of the sedimentary pile. Similarly, Korn and Martin (1959) have described nappe-like structures with detachment planes up to 20 miles in length and movements of up to 3000 feet of (non-geosynclinal) sediments over distances of 10 miles. This was attributed to upwelling on the edge of the basin causing free gliding toward the centre.

In the Gulf Coast of Louisiana, Bornhauser (1958) described large folds due to mass "sedimentary creep" of large thicknesses down a 2° or 3° slope. The anticlines are narrow and elongate, with shallow dips on the flanks. They are up to 20 miles in length, and have axes parallel with the regional strike. The beds show a thinning on the crests of anticlines and a corresponding thickening down the flanks. Normal faults are associated with the folding. They dip at 45° , and have a "down-to-the-coast" movement. The faulting is considered to be a near-surface expression of the deeper seated folding.

Finally, there are the great slices of allochthonous Argille scagliosa which slid off the crests

of a series of rising submarine ridges, and travelled for distances of 80 miles into the fore-deep of the Northern Appenine trough (Maxwell 1959).

These examples of gravity movements can be thought of as tectonic, or more specifically, early tectonic. They are distinct in concept from the sedimentary slumps because they are of large scale, they constitute bulk deformation of the sedimentary pile, and the deformation is slow, coherent and continuous.

There is one type of soft-sediment deformation that is difficult to classify according to these criteria. This is the down-slope gravitational sliding of thousands of feet of sediment by mass translation without crumpling. Baldry (1938) and Brown (1938) described sheets of breccias and contorted bedding in the Tertiary of Ecuador and Peru which were interpreted as deep-seated, basal slide planes for such a process. It should be noted however, that doubt has been cast upon the interpretations of Baldry and Brown by Marchant and Black (1960) who believed that apart from minor faulting, all structures are open cast and penecontemporaneous resulting from submarine slumping, possibly load casting and convolute folding. The process of gravitational sliding as envisaged by Brown (1938) should probably be grouped with the early tectonic processes since it constitutes bulk deformation of the sedimentary pile.

From these few examples, it is maintained that there is an important difference in concept between sedimentary and tectonic structures. The distinction is mainly on scale and cause of deformation. Scale alone is not sufficient, since the distinction would then be an arbitrary one. The two types of structures are defined below.

Sedimentary structures are those which form in unconsolidated sediments, at or very near the sediment-water interface, by the action of water-current stresses, hydrostatic stresses or gravitational stresses, and are unrelated to bulk deformation of the sedimentary pile. Where a sedimentary pile consists of an alternation of disturbed and undisturbed layers, and if it can be shown that the disturbance was an open cast deformation, this does not constitute bulk deformation of the sedimentary pile.

Tectonic structures are those large-scale structures, which may occur in consolidated or even unconsolidated rocks, formed during bulk deformation of the sedimentary pile. The immediate cause of deformation may be gravity sliding but the ultimate cause is vertical movement.

THE CRITERIA

1. External form

Sedimentary deformations are generally confined

to thin, laterally persistent horizons which may contain one or more sedimentation units. The bedding separating these deformed zones is undisturbed.

2. Establishment of contemporaneous erosion

Where the upper boundary of a disturbed zone truncates the structures and is a plane of erosion, penecontemporaneity is established and the disturbance is sedimentary. The plane of erosion may be proved by (a) the presence of cross-lamination infillings in depressions, (b) the presence of a thin graded bed resting on the truncation plane, (c) the presence of load casts of the overlying undisturbed material in the disturbed material, and (d) pieces of deformed underlying rock in the overlying rock.

3. Style

Sedimentary contortions tend to have an overall style indicative of flowage, although this in itself is not sufficient to prove contemporaneity. There is one aspect of style which enables sedimentary folds to be distinguished from tectonic folds. The joint occurrence of rounded synclines and sharp cusped anticlines across which the bedding laminations are thinned, is a good criterion of the types of sedimentary folds due to vertical movement. This feature has been remarked on by many writers, for example, Sanders (1960), Dott and

Howard (1962), Selley et. al. (1963), Stewart (1963), Keane (1958), Sutton and Watson (1960), and is well illustrated in structures in the Rocky Cape Group. This phenomenon should be clearly distinguished from similar structures formed during concentric folding involving lateral movement along a basal decollement and without thinning over the anticlines.

4. Associated structures

The most frequently cited feature of sedimentary folds is the absence of slickensides, cleavage, veining, and crushing in rocks that are extremely contorted and brecciated. This is true only in so far as it establishes that the rocks were unconsolidated at the time of deformation and other criteria must be used to distinguish early tectonic from sedimentary deformation. Under some circumstances, a flow cleavage is alleged to form (Jones 1940, p.341) by streaking-out of small particles and disrupted fragments of the bedding by laminar viscous flow. The product is stated to be a rock resembling an Archean gneiss.

There is a distinct source of confusion where penecontemporaneously deformed sediment is subject to later tectonic folding and cleavage formation. Some processes of slumping and load deformation break the layer into pillow-shaped detached fold cores of sand

separated by pelitic material. This marked anisotropism would influence the location and orientation of later cleavage, especially where the limbs of isoclinal convolutions are oriented at a favourable angle to the bedding. The result is a minor fold having an axial-plane cleavage. This has been observed in the Detention Sub-group in the Rocky Cape Group and illustrated in Figure 2. Similarly, load casted flute marks in the Burnie Formation, which on any one bed have a parallel orientation, also control the cleavage.

5. Orientation

Another criterion, is that sedimentary folds tend to have a haphazard orientation. Those structures formed by vertical movements have axes with random azimuths and variable plunges. Slump folds generally have axes lying parallel to the bedding plane and may even have a high degree of preferred orientation due to the directional movement down the regional slope. This is the case in the interlayered siltstone in the orthoquartzites of the Rocky Cape Group, but not so in the slump folds in the Burnie Formation.

The orientation of sedimentary folds is often said to bear no relation to the orientation of the major tectonic axes. (Nevin 1949, p.202). This may be true for some convolute folds, but slump folds, which are directional structures, tend to have their axes aligned along the palaeostrike of the basin. In so far as the

main folding of a tectonic belt is parallel to its elongation, it is not exactly coincidental if slump fold axes are parallel to the tectonic axes. This regional relationship is a gross over-simplification and would be so vague geometrically as to be only statistically observable.. This contrasts markedly with the precise correlation between orientation of coeval minor and major structures.



Plate 14a Pseudo-nodules of sandstone in
siltstone, 300 yards east of the
headland at Jacobs Boat Harbour.



Plate 14b Diapiric contortion in ortho-
quartzite, headland at Jacobs Boat
Harbour.

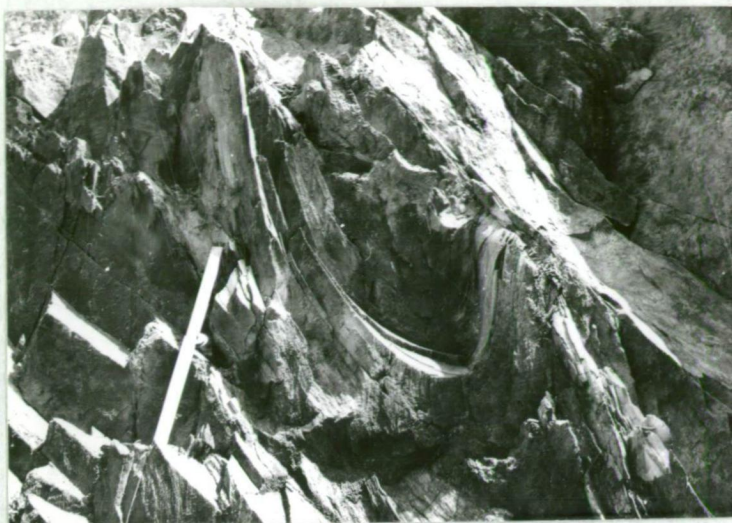


Plate 15a Rounded syncline and cusped anticline to the right, in zone of ball-and-pillow structures, Detention Sub-group, railway cutting, Sisters Hill.

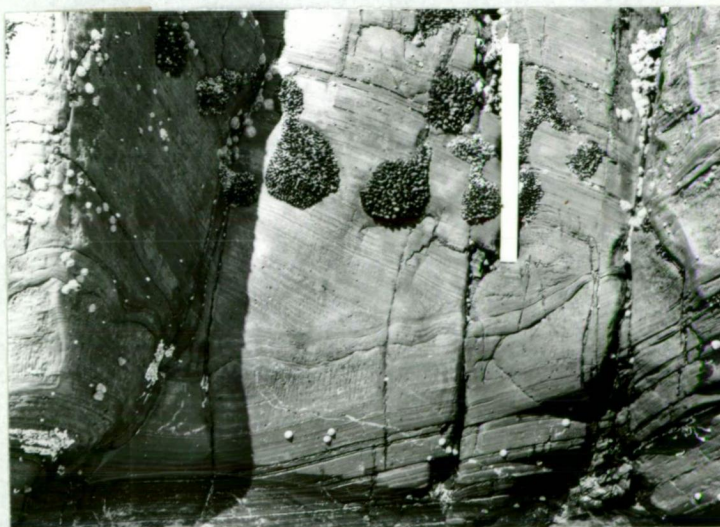


Plate 15b Detail of basal plane of block slide, Cowrie Siltstone, Rocky Cape.



Plate 16a Slump sheet in Burnie Formation,
western end of West Beach, Burnie.



Plate 16b Slump sheet in Burnie Formation,
Round Hill Point.



Plate 17a Sedimentary schuppen structure in cross-bedded sandstone lenses in shale, Cowrie Siltstone, Cowrie Point.

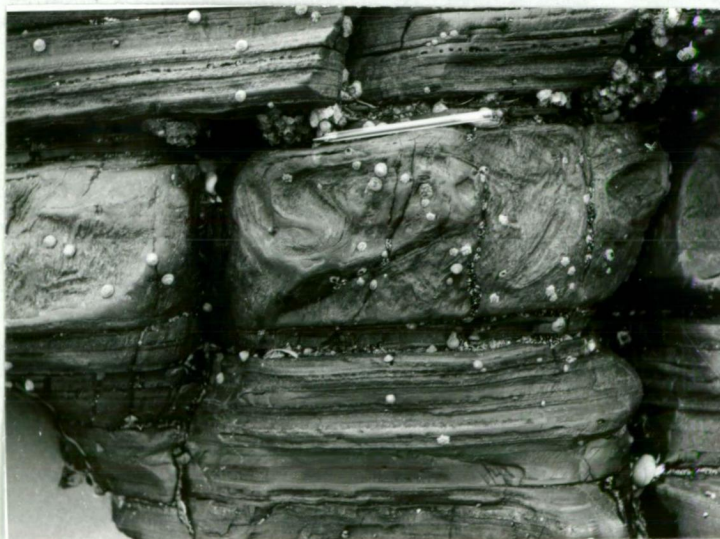


Plate 17b Convolute lamination in fine sandstone within shale, showing the contortions dying out against top contact, Cowrie Siltstone, west side of Rocky Cape.

C H A P T E R 6

SEDIMENTARY HISTORY OF THE PROTEROZOIC BASINSEDIMENTATION IN THE ROCKY CAPE GROUPEarly starved basin

The earliest record of sedimentation in the Proterozoic basin is given by the lower part of the Cowrie Siltstone. This is a fine-grained shale with a conspicuous and delicate lamination. Detritus coarser than medium silt is rare. Evidence of bottom current activity is given only by an extremely fine wavy lamination within the silty layers. Graded bedding and exotic pebbles are absent. The features suggest that deposition occurred in a quiet, aqueous environment, in water deep enough where only clay and fine silt accumulated by fallout from suspension in slowly flowing currents.

As sedimentation proceeded, small incursions of silt-laden currents appeared in the otherwise quiet basin. The depositional structures in the silty beds, such as sets of festoon cross-beds, discrete cross-bedded lenses and ripple marks, indicate a variable and intermittent supply of the coarser material. This also indicates an increase in current competence and suggests a shallowing of the basin.

The orientation of ripples and cross-bedded lenses, together with the diminution of grain size from east to west, indicate that the currents were coming from the southeast. Despite these minor incursions, the environment was basically starved for detrital material and strongly reducing, with mainly fine-grained laminated black pyritic shale accumulating.

After the accumulation of at least 8,000 feet of sediment, the environment changed from a quiet, euxinic and starved basin, to one of shallow water, free-circulation and high energy. This change may be correlated partly with a filling of the basin of deposition, and partly with an increasing supply of sediment. Thus there was an excess of sedimentation, although slow, over an even slower subsidence.

Environment of the orthoquartzites

The depositional environment of the pure quartz sands is thought to be shallow water, probably marine, in either the sub-littoral or neritic zones. Super-mature chemical composition, high sphericity, good rounding and size-sorting attest to the high energy of the environment. Multiple sets of medium-to large - scale cross-bedding and small-scale ripple mark are common. The bedding proper is planar and regular. These features, plus the interbedded lutite rule out an aeolian or fluvian environment.

The thick accumulation (total of 9,000 feet) of orthoquartzitic sand indicates a remarkably constant physical environment. This long standing stability was interrupted only when the interbedded lutites were deposited.

The thickest of these lutite beds (Irby Siltstone) consists of a sequence of: a black pyritic shale with discrete cross-bedded lenses similar to parts of the Cowrie Siltstone; a bedded clastic dolomite; a horizon of interbedded orthoquartzite, siltstone and mudstone with abundant shallow water depositional structures; and a sub-graywacke horizon with detrital hematite.

This temporary return to the quiet starved euxinic conditions may be interpreted by temporary bar-restrictions across the inlets of sedimentation or by a sudden deepening of the basin. With progressive filling of the local basin, the environment changed to stable shallow water with gentle currents and deposition of dolomitic mud, then to a littoral oxidising environment, and then back to the stable sub-littoral environment.

The vertical passage of the orthoquartzite into the Port Slate is expressed as a thinning of the bedding units and the appearance of interbedded siltstone. Eventually the sandstone forms trains of discrete cross-bedded lenses. This also represents a temporary return to the starved basin conditions of parts of the Cowrie Siltstone.

Palaeocurrents and Palaeogeography

It is possible to reconstruct in broad terms the palaeogeography during deposition of the Rocky Cape Group, by considering mainly palaeocurrents and lithological associations. Because of lack of information the numerous palaeo^{geo}graphic indicators such as facies changes, sand-shale ratios, pebble size, isopachs, and low-angle unconformities cannot be successfully applied, and hence a fully integrated palaeo^{geo}graphic reconstruction cannot be made.

Palaeo^{geo}graphic reconstruction is difficult because cross-bedding directions in marine sands must be considered in relation to other palaeocurrent indicators. This arises because current directions in the littoral and neritic environments are redispersal currents induced by tidal currents, prevailing wind drift or estuary outflow, and are not the primary transporting currents bringing sediment from source to basin. Whereas in fluvial and deltaic environments, there is good reason to believe the mean cross-bedding direction parallels the dip of the palaeoslope, there is no a priori reason why this is so in marine sands. Consequently, there is no guarantee of a simple relation between the transport direction and the orientation of the strand line.

Some basis for correlation of the cross-bedding direction with the palaeoslope may be given by comparison with similar sandstones where the facies variations are accurately known. In the Lower Palaeozoic of the Upper Mississippi area, Farkas (1960), Hamblin (1968) and Potter and Pryor (1963), found that the relationships between cross-bedding, facies, unconformities and fauna all point to the palaeoslope being equated to the mean cross-bedding direction. Pettijohn (1957) also believed that the cross-bedding direction in the thick Lake Superior Precambrian quartzites was indicative of the palaeoslope because of the remarkable persistency through long periods of geological time. Tanner (1956) in the cross-bedded Pleistocene oolite off the coast of Florida found a strong mode indicating an off-shore current, and two other modes at right angles, indicating long-shore currents. Palaeogeography in this case was deduced from present day geography.

The concept of the dominant current direction being parallel to the palaeoslope, and hence at right angles to the palaeostrike and strand line is supported by the following evidence in the Rocky Cape Group orthoquartzites.

- (a) The persistence in current direction over prolonged period of geologic time indicates an overall tectonic control rather than control by

local strand line conditions as would be the case with long-shore currents.

- (b) The evidence given by the inclination of cross-beds indicates that the palaeoslope has a 1° - 2° component in a northwesterly direction.
- (c) Both the maximum and the mean thickness of cross-bedding sets are less in the west than they are in the east. Pelletier (1958) and Schwarzscher (1953) found such a decrease in size down the palaeoslope.
- (d) The vector-mean of ripple axes in the Detention quartzite trends north-north-east, and the minimum direction corresponds to the cross-bedding direction. There is good evidence (Potter and Pettijohn 1963, p.94-98) that ripple axes in marine shelf sands and littoral sands will outline the depositional strike.
- (e) The direction of slumping in the interbedded lutites is parallel to the dominant cross-bedding mode in the immediate vicinity.

All the palaeocurrent patterns in the Detention quartzites, (Figure 4), show a strong mode tightly grouped to the northwest, with a weaker, diametrically opposed southeasterly mode which shows a little more variance. With the exception of one station in the Jacob Quartzite,

the northwest mode is always present. It is therefore concluded that the regional palaeoslope dipped gently to the northwest, and the strand line lay somewhere to the southeast.

The bimodal pattern occurs repetitiously throughout the basin and must have some significance. Such polymodal palaeocurrent patterns appear to be uncommon in the geological literature. The cross-bedding is believed to be remnants of transverse mega-ripples formed on the shallow sea floor under the influence of steady currents. If the cross-bedding was the "herring-bone" type in which succeeding sets are oppositely directed, the bimodal pattern may be explained by flood and ebb-tide currents moving backward and forward in a direction perpendicular to the strand line. However, the sampling procedure reveals that it is not of the "herring-bone" type, but rather a clustered type in which azimuths of one mode are stacked continuously one upon the other.

If the dominant northwest mode is directed down the palaeoslope, then the subsidiary, diametrically opposed mode must also represent a transient sloping of the palaeosurface to the southeast which at times was dominant over the regional palaeoslope. The subsidiary mode would have an associated strand line somewhere to the northwest. It thus appears that transverse dispersal currents were operative in this basin, which,

on extrapolation from the regional trend of the Proterozoic basin, was oriented northeast-southwest. In the environment envisaged, namely a broad flat sandy plane covered by shallow water (less than wave base), these reversals in palaeoslope could be achieved by slight differential subsidence. This in turn suggests a flanking fore-land to the northwest as well as the older basement to the southeast. The major structural evidence also supports the idea of a basement foreland to the northwest, (Chapter 9). The relative strength of the respective modes then gives some indication of the relative stability of the flanking forelands.

The tectonic significance of thick blanket orthoquartzite is uncertain and even the thinner blanket sands present problems. A phenomenal total thickness of 8,500 feet of mature quartz sands was deposited in this basin. Well-sorted mature sandstones imply a tectonically stable environment with considerable wave and current action to winnow the finer material and round the residual quartz grains. Extensive blanket orthoquartzites are generally interpreted as continental transgressions of the shoreline. This interpretation is hardly applicable because of the extreme thickness involved. It would appear that the sands were laid down in a broad flat basin undergoing slow continued subsidence and maintaining a delicate balance with the rate of deposition,

so that the position of the strand line and the depth of water remained approximately constant. The balance between accumulation and subsidence would be aided by continual dispersal of sand across the axis of the basin under the influence of the transverse currents. In this way the sand was probably zig-zagging its way along the axis of the basin and becoming trapped in smaller sub-basins of subsidence.

This environment is usually thought of as "stable shelf", however these orthoquartzites belong to a thick sequence which may be considered thick basinal or geosynclinal, rather than stable shelf. A thickness of 8,500 feet of sediment also demands long, gentle and continuous subsidence which is not altogether consistent with the concept of tectonic stability. This contradiction may be partly resolved by postulating very long transportation of detritus that might offset the effects of large geosynclinal subsidence which normally produces immature sediments. However, tectonic stability over very long periods, perhaps as much as 100 million years appears necessary.

Provenance of the orthoquartzites

The palaeocurrent indicators show that the dispersal currents were coming mainly from the southeast where there is an older crystalline basement (Tyennan Nucleus) composed of metasedimentary quartzites and

schists. However it does not appear likely that the material in the Proterozoic basin was shed from the Tyennan Nucleus and transported to the northwest. As stated previously, in marine sands, without an accurate knowledge of the palaeogeography, the palaeocurrent directions give no clue to the direction of the source rock.

The texture and mineralogy of the orthoquartzites rule out a source from the central Tyennan Nucleus of Tasmania. About 99% of the grains are single-crystal quartz grains of medium to coarse-sand size, which is much coarser than even the coarsely crystalline metaquartzites. This indicates that the source rock was primarily granitic or possible gneissic. This is supported by the presence of some rare well-rounded grains of orthoclase and microcline in the Detention quartzite. Polycrystalline quartzite, chert fragments, rounded green tourmaline and zircon are present in very small quantities. This assemblage may indicate that the orthoquartzites are, in part, recycled sediments. No textural evidence in the form of multiple secondary overgrowths on the quartz grains are present to confirm the hypothesis of recycling.

Potter and Pryor (1963, p.1227) believe that mature sands cannot be produced in large quantities from crystalline basement rocks by the normal processes of erosion, transport and deposition; and maintain that such sediments demand recycling. They cite the St. Peter Sandstone of Ordovician age (Upper Mississippi Basin) that is derived by the double recycling of detritus shed from the Canadian Shield. Krumbein and Sloss (1963, p.549) on the other hand, consider that such sediments can be produced by long continued chemical and mechanical maturation of grains under stable tectonic conditions in a single cycle. However, because of the large thicknesses involved, recycling was probably an important factor.

It is concluded that the sands were derived from a distant granitic or gneissic source of unknown locality, and have been transported, deposited and continually reworked in shallow water by currents that were transverse to the axis of the basin.

INITIAL EMERGENCE OF THE ROCKY CAPE GEANTICLINE

What may be taken as the first stage in the rise of the Rocky Cape Geanticline is recorded by a gentle subsidence in the west and a more pronounced subsidence in the east. On the western side there accumulated a stable fore-land facies consisting of an orthoquartzite-dolomite suite (Smithton Dolomite),

whereas on the eastern side there was deposited a thick sequence of quartz wacke having mild eugeosynclinal affinities (Burnie Formation). Since both these contrasted suites were derived in part from the older Rocky Cape Group, this indicates a slight emergence of the geanticline. The main stages before and after this initial emergence are shown in Figure 12.

The age relationship between these two suites is uncertain, as they are not in contact. Both assemblages are younger than the Rocky Cape Group and both are unconformably overlain by Cambrian volcanics, and it appears that they are broadly contemporaneous, each assemblage indicative of contrasted tectonic environments on the flanks of the newly emergent geanticline at approximately the same time.

Smithton Basin

It is proposed to call the basin that formed on the western side of the newly emergent ridge, the Smithton Basin. The shape of the Smithton Basin is more or less outlined by the present known limits of the Smithton Dolomite which is an extensive, regionally transgressive blanket about 1000 feet thick. This blanket extends from the Black River across the axis of the trough to Marrawah, 40 miles west of the mapped area; and extends for a considerable distance to the south along the inferred axis of the basin. The dolomite is finely crystalline and well laminated, with minor detrital

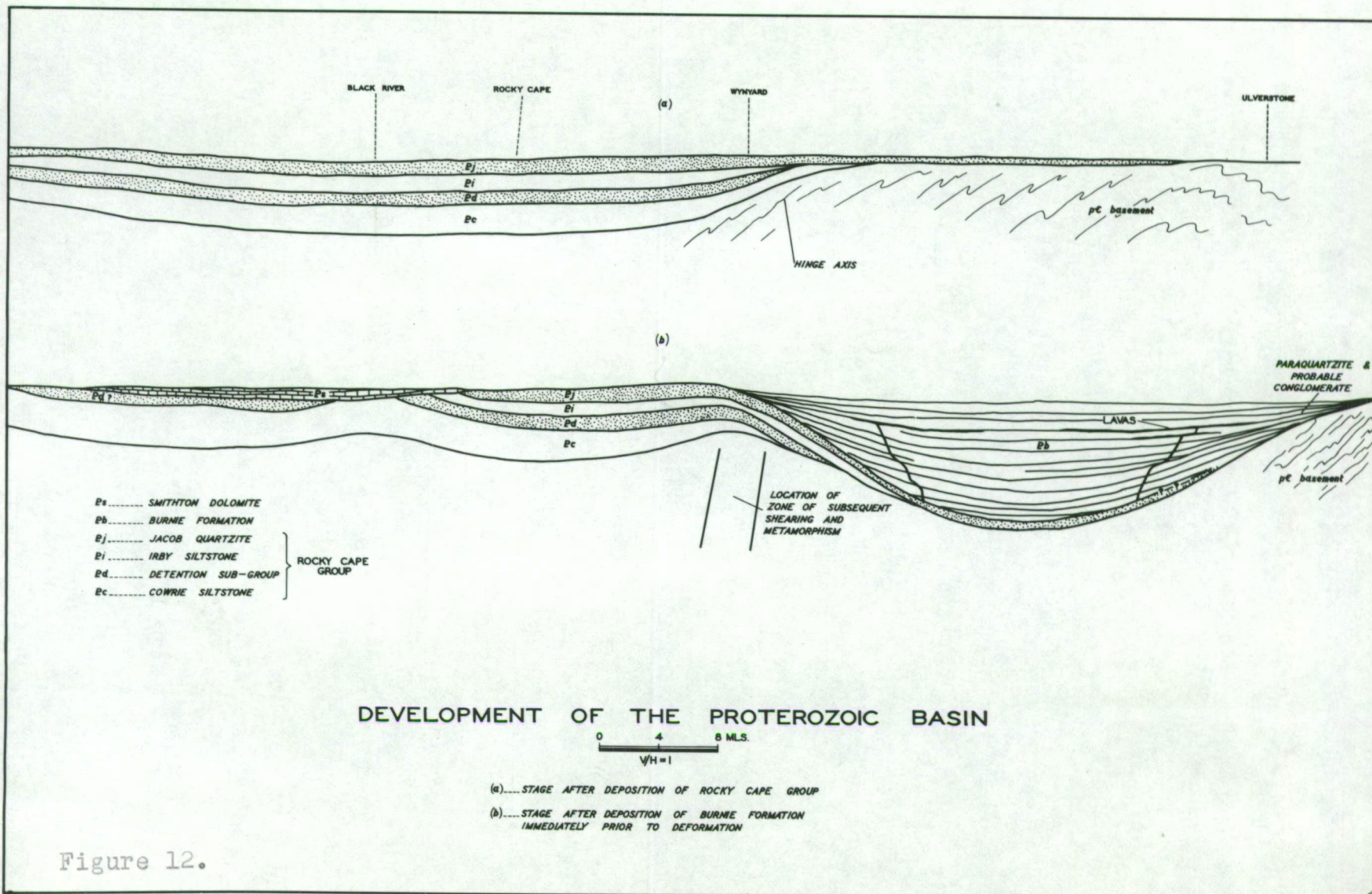


Figure 12.

quartz grains. The dolomite was precipitated on a broad shallow basin having extreme tectonic stability, into which little detrital material passed.

On the ^{ea}western edge of the Smithton Basin at Black River, the dolomite overlies the Cowrie Siltstone which is low down in the Rocky Cape Group. This probably was approximately where the initial emergence occurred, off which there must have been a considerable amount of erosion (Figure 12-b). As subsidence continued the axis of the newly emergent ridge shifted slightly to the east to about the present position of Rocky Cape.

In the Black River area the dolomite has a basal horizon 70 feet thick, of clean orthoquartzitic conglomerate and sandstone. This is probably a shore-line deposit, which transgressed onto the newly emergent ridge as its axis shifted to the east. The conglomerate and sandstone was derived from the orthoquartzite of the Rocky Cape Group which stratigraphically overlies the Cowrie Siltstone further to the east. These quartzites probably formed the main emergent ridge.

The significance of the 18° angular discordance between the basal conglomerate and the Cowrie Siltstone is uncertain since shore-line conglomerates may have initial dips of up to 20° . The angular discordance therefore does not necessarily imply folding prior to deposition in the Smithton Basin, and it is clear that

the major folding in the Rocky Cape Group (Penguin Orogeny) post-dates the dolomite deposition.

The present whereabouts of the bulk of the material that was shed from the newly emergent ridge is uncertain. Some of the material went into the Smithton Basin, and some was incorporated into the Burnie Formation. It is possible that the bulk is represented by the Donaldson Group of Spry (1964) which is an orthoquartzitic sandstone and conglomerate unit, occurring about 50 miles to the southwest along the axis of the main Proterozoic basin.

SEDIMENTATION IN THE BURNIE FORMATION

Basin of Deposition

The Burnie Formation is a thick flysch-type deposit, which accumulated in a basin between the newly emergent geanticlinal ridge to the west, and the older Precambrian crystalline basement to the east. Little is known about the size and shape of the basin of deposition for reasons such as the absence of marker horizons, the structural complexity, the lack of inland outcrop and the nature of the study which is essentially a cross-sectional investigation. Stratigraphic subdivision is not possible, and the thickness variations of the formation as a whole are not known. These problems are especially common to the flysch-type basins generally, which usually are the loci of more intense tectonic

deformation and are thus usually sheared from their bases.

The structural profiles indicate that there is a minimum thickness of 14,500 feet, but neither the base nor the top is exposed, and the true thickness could twice this amount.

Spry (1964) has suggested that the Oonah Quartzite in the Zeehan area is equivalent to the Burnie Formation. This is supported by the direct correlation of the Keith Metamorphics with the Whyte Schist, and also by the regional northeast-southwest fold trend. These points are discussed further in Chapter 12. In broad terms, the flysch-type sediment can be thought of as occupying a northeast-southwest linear trough, flanked to the east by the older Precambrian basement. After removing the effects of lateral shortening from the structural profiles, this trough is seen to have an initial width of 50 miles.

The Burnie Formation, although younger, does not directly overlie the Rocky Cape Group, but onlaps it so that the axis of greatest sedimentation is offset to the east. The reconstruction of the sedimentary pile at this stage is shown in Figure 12-b. The zone now occupied by the Keith Metamorphics appears to have acted as a hinge line, around which subsidence first occurred for Rocky Cape Group sedimentation and later for the Burnie Formation sedimentation.

Thus a remnant basement high is shown in Figure 12-b. This high is in no way similar to the median ridge of Kay (1951) which is of volcanic origin and separates concurrently depositing miogeosynclinal and eugeosynclinal facies. It is non-volcanic and does not contribute in any way to the supply of sediment for either basins.

Agents of Sedimentation

The lithologic features that may be significant in deducing the depositional history in terms of water depth, deposition medium, and current activity are (1) the rhythmical bedding, (2) the lithological association of black lutite and quartz wacke, (3) the high proportion of matrix, (4) the lamination and cross-lamination within the arenite, and (5) the sharp basal contact and the gradational top contact of the arenite beds.

Some of these factors are often taken as indicative of turbidity current activity. However, the most characteristic feature of turbidites, i.e. repetitious grading is not present. Thus the process first suggested by K  nen and Miglorini (1950) of flash incursions of silty detritus into a standing body of water producing a graded unit without lamination is not realistic in this case. At a later date K  nen and Menard (1952) explained the lack of grading in the body of the turbidite unit by a continual feeding of the detritus.

Although adequately accounting for the type of grading observed in the Burnie Formation, this idea does not explain the lamination, cross-lamination and ripple marking. The concept of the turbidity current was considerably broadened after Känen (1953) invoked the concept of the dilute tail behind a waning turbidity current to account for the structures that are obviously the result of traction transport along the bottom.

Consequently, many graywacke deposits with internal lamination and cross-lamination were interpreted as turbidite sequences, for example, Van Houten (1954), Dzulynski and Radomski (1955), Ballance (1964), Wood and Smith (1958), Bouma (1962), and many others. Thus it has come to be accepted that there is a typical turbidite sedimentation unit in which a specific sequence is found. This sequence consists of: at the bottom, a coarser graded massive interval resting on an erosional surface; a middle interval of parallel lamination; an interval of small-scale cross-lamination which is frequently convoluted; and an upper layer of parallel lamination which grades rapidly into the overlying lutite horizon. Bouma (1962) also showed that when this sequence was incomplete it was usually the basal intervals that were absent.

As has been outlined in a previous chapter, this association of structures, as well as the basal cut-outs, generally is present in the thinner arenite beds. The thicker arenite beds have a sequence commencing with a sharp basal contact on an erosional plane, followed by an interval of thin lamination, followed by a thicker interval of festoon cross-bedding, then an interval of parallel lamination passing quickly up into the lutite. The planar lamination is a fine layering and is not the expression of repetitious grading due to the passage of a series of turbidity currents, but rather deposition layer by layer by traction movement of independently behaving detrital grains.

Festoon cross-bedding of this size is not usually developed in turbidite deposits. Normally, the thickness of the individual cross-bedding units averages one or two inches, and this type should be differentiated from the larger festoon cross-bedding. The cross-bedding structures illustrate the two natural classes of size recognised by Allen (1963b).

The presence of two petrographically similar but lithogenetically different types of arenites, namely the turbidite and traction deposits raises the question of how the traction deposits contain so much matrix. It appears that in the traction deposits much of the

fine clay matrix was deposited along with the fine sand and coarse silt. Lombard (1963) used the term laminite for flysch-type rocks which are internally laminated containing abundant matrix. Lombard thought these to be deposited from a cloud of muddy suspensoids that settle afterwards or in front of the cloud, rather than a turbid mass of sediment that slides quickly along the floor. However, according to Hjulstrom (1938) it is not possible to deposit from suspension the fines with the coarse fraction moving in traction. It is suggested that if the sand silt and clay falling from suspension, together with the sand brought in by bottom currents are deposited concurrently, the normal process of size sorting may not be permitted if the bottom currents are transient and non-directional. There is evidence from the festoon cross-bedding described in a previous chapter that these bottom currents were variable.

The type of environment envisaged is a dominant regime of deep, quiet water in which black, faintly laminated mudstone was deposited. Some of the mudstones contain pyrite cubes pointing to a basin with poor bottom circulation. At frequent and regular intervals this quiescence was interrupted by the incursion of dense currents laden with sand, silt and clay. Deposition occurred mainly by dropping from suspension with secondary currents being set up as the result of the passage of the density current. During and after deposition, the

sediments were reworked and added to, by ocean bottom currents of considerable magnitude, quite independent of those currents set up as the result of the passage of the turbidity currents.

The depth of water under which the coarser sediments were deposited and reworked is difficult to evaluate. The morphology and scale of cross-bedding may provide some clue by analogy with modern sediments. Cross-bedding of this type probably results from the migration of linguoid ripples. Recent experimental work by Allen (1963a) suggests that some proportional relationship exists between amplitude of ripples and depth of water, but much more work is needed. Estimates of the maximum depth are very unreliable in the light of recent oceanographic discoveries of ocean bottom currents at great depth on the abyssal plain. This has led Hubert (1964) to suggest that much coarse silt and fine sand found on the abyssal plain may be turbidites reworked by ocean bottom currents.

The presence of black mudstones, the association of pillow lavas, the absence of typical shallow water structures and the regularity of bedding together suggest that the sediments were deposited in fairly deep water.

Provenance

The problem of position and nature of the source rocks for the Burnie Formation is similar to that of the

Rocky Cape Group. Mineralogically, the two groups are similar, the arenite being composed of single-crystal quartz grains. Polycrystalline metamorphic quartzite grains are generally absent, so that derivation from the older Precambrian crystalline basement was not an important factor. Some older metamorphic material may have been incorporated into the cleaner arenite at Penguin which contain minor amounts of polycrystalline quartz.

Much of the quartz wacke, although texturally immature, contains what are clearly recycled grains. These are rounded single-crystal quartz grains with a secondary overgrowth which is often fragmented. This may indicate some derivation from the rocks to the west. The presence of accessory tourmaline which is well-rounded and occasionally fragmented also supports this idea.

Palaeocurrent indicators from the Burnie Formation suggest a dual system of sediment dispersal. Flute marks indicate axial sedimentation by turbidity currents in a northeast-southwest direction along the depositional trough. The bulk of the material in the Burnie Formation may have been brought in by these axial currents from a distant and unknown source of dominantly sedimentary terrain. Festoon cross-bedding (linguoid ripples) and transverse ripples indicate a direction of sediment movement transverse to the axis of the trough.

These transverse currents reworked the sediment previously deposited by turbidity currents, and may also have introduced new sediment into the trough from the newly emergent ridge to the west. The presence of transverse currents is also suggested by a marginal change in the east at Penguin from quartz wacke to sub-graywacke and protoquartzite. The quartz conglomerate noted by Burns (1964, p.27) at Goat Island near Ulverstone may also be the expression of such a facies change.

C H A P T E R 7

COOEE DOLERITE

The presence of dolerite bodies in the Precambrian sediments on the northwest coast was first recorded by Twelvetreets (1903), but were first described in detail by Spry (1957a). The dolerite bodies outcrop sporadically along the coast from Sulphur Creek to Black River with the greatest concentration between Burnie and Cooe. Consequently, Spry (1957a) termed this group of rocks as a whole the Coee Dolerite. Structural and chemical evidence was presented suggesting that these intrusions were Precambrian in age, rather than feeders for the Cambrian spilitic lavas. This idea found support in a single radiometric K/A dating of 700 million years (Spry, 1962). Later (Spry, 1964, p.48) suggested that the dykes in the lower Pieman River area were equivalent to the Coee Dolerite.

The present study adds little to the petrological considerations of Spry (1957a), but the structural relations have been examined in some detail.

PETROLOGY

The intrusives may be described as deuterically altered sodic dolerite. Petrographic variation is due to different degrees of deuteric alteration and shearing. Several primary minerals are recognisable in thin section.

Pyroxene occurs in relatively fresh crystals

up to 2mm showing border alteration to pale fibrous amphibole and chlorite. It varies in colour from pale green to pale pleochroic purple. The optic angle varies from 45° to 66° and Spry (1957a) records an analysis showing it to be high in Al_2O_3 (9.0%), and TiO_2 (1.60%) and low in CaO (17.7%).

Small quantities of sericitized orthoclase are present in the strongly altered background. Fresh orthoclase occurs in graphic intergrowth with quartz in dolerite from Detention River (33358) and Edgumbe Beach (33357).

Hornblende, in subhedral prisms up to 1.5mm, occurs in the coarse dolerite at Parklands and Coee Point. It is commonly found surrounding the pyroxene and terminated against the euhedral faces of the pyroxene. It is strongly coloured with X- pale green, Y - reddish green, and Z - deep green or rusty brown, and has a maximum extinction $Z \wedge C$ of 13° .

Spry considered the plagioclase in the rock to be secondary albite derived by alteration of a more calcic plagioclase. Some of the leucocratic varieties from Parklands (eg. 4851, 4852) contain subhedral, compositionally-zoned crystals which suggests that they are remnant crystals of primary plagioclase. These were investigated by the universal stage method of Slemmons (1962). The crystals are twinned on both the

Albite Law and the Albite-Carlsbad Law, and seven determinations gave a range between An 44% and An 63%. Four determinations in 4851 gave 57% An, indicating labradorite. This confirms that the albite is not a primary mineral.

Minor amounts of well-formed pleochroic (yellow-brown) biotite up to 1.0mm occur in the dolerite from Parklands. Quartz is a common but subordinate constituent, occurring interstitially to the albite, or in graphic intergrowths with orthoclase. Patches of serpentine having the euhedral form of olivine are probably pseudomorphs after olivine. Titanium minerals are ubiquitous accessories. Ilmenite occurs as tabular flakes, commonly in aggregates which interlock in a triangular network. Quartz, sphene magnetite and leucoxene are often associated. Accessory apatite occurs in needles up to 3.0mm.

A variety of secondary minerals are the product of the albitization (saussuritization) and uralitization.

Albite generally occurs as laths and also as shapeless grains. It is abundantly included with zoisite with lesser quantities of chlorite, sericite and sphene. The albite and zoisite are derived by saussuritization of labradorite. Small amounts of prehnite and calcite, interstitial to the albite, are also produced.

Pale green actinolite occurs as a fibrous fringe around, and penetrating, the pyroxene. Complete replacement of pyroxene by actinolite is found. The amphibole shows all stages of alteration to a colourless chlorite with a low birefringence, often with anomalous blue interference colours. The same chlorite also occurs scattered throughout the heavily altered fine-grained background between pyroxene and feldspar crystals.

All the dolerite bodies are mineralogically and chemically related, and are considered to be the result of the one period of igneous activity. Five chemical analyses are listed in Table 4, three of Spry (1957a) and two new ones. The table shows the broad similarities although there is a considerable range in the $\text{Na}_2\text{O}:\text{K}_2\text{O}$ ratio of 0.7 to 2.6. This index is often used in attempting to correlate lower Palaeozoic intrusives in Tasmania, (Spry 1957a, Urquhart 1966) but it appears to be too variable to be useful.

Spry (1957a) divided the dolerite into five types using the criterion of degree of alteration. On this basis there are two main types which have some structural significance.

(a) moderately altered dolerite

The moderately altered dolerite generally occurs in the larger tubular bodies greater than 10-20 feet thick.

TABLE 4

ANALYSES OF DOLERITE

	1	2	3	4	5
SiO ₂	44.44	58.32	54.64	54.0	48.5
Al ₂ O ₃	17.97	15.03	15.22	14.0	11.5
Fe ₂ O ₃	1.74	0.96	1.17	0.5	0.7
FeO	8.32	2.60	7.81	10.0	8.3
MgO	4.74	4.36	5.39	5.1	4.4
CaO	9.42	8.46	7.96	8.3	6.3
Na ₂ O	2.86	3.24	1.69	1.9	2.6
K ₂ O	3.04	1.41	2.23	1.9	1.0
H ₂ O-	0.26	0.14	0.02	2.6	4.0
H ₂ O+)	3.53	5.04	2.85	0.26	0.21
CO ₂)				0.10	-
TiO ₂	3.00	0.41	0.68	0.74	1.2
P ₂ O ₅	0.72	0.12	0.12	0.10	0.16
MnO	0.22	0.11	0.13	0.17	0.16
	100.26	100.20	99.89	100.06	99.33

- 1 Coarse dolerite, Burnie, specimen 4864. (Spry, 1957a)
- 2 Dolerite, Detention River, 4852. (Spry, 1957a)
- 3 Leucocratic dolerite, Burnie, 4861 (Spry, 1957a)
- 4 Coarse dolerite, Detention River, 33358 (new analysis)
- 5 Sheared dolerite, The Port, Rocky Cape, 33359
(new analysis)

Analyst for 4 and 5: H. K. Wellington.

These are hard massive dark-green medium-grained rocks, and occur at Sulphur Creek, west of Blythe Heads, between Burnie and Cooee, Detention River and Edgcumbe Beach.

In thin section, the dolerite texture is still recognisable by randomly oriented interlocking laths of feldspar with some interstitial augite. The doleritic texture is partly destroyed by alteration to give a mosaic of shapeless interlocking albite grains. This mosaic texture closely resembles the texture of some of the basic schist in the Keith Metamorphics, some of which display remnant doleritic textures, (Chapter 8).

Some minor varieties of dolerite are associated with the main body at Parklands. Micro-dolerite occurs in the chilled margin of some of the small dykes and irregular offshoots of the main body. These consist of a fine mesh of albite needles with small ragged biotite flakes in a chloritic background.

Coarse-grained pegmatitic dolerite occurs in irregular patches. It contains large titanite (up to 3mm) with ophitic albite laths and is surrounded by fresh subhedral hornblende. Minor interstitial quartz is present with long included needles of apatite.

Leucocratic dolerite occurs as dykes within the main body at Parklands. This consists of zoned euhedral to subhedral labradorite and anhedral albite, with lesser quantities of skeletal tremolite, quartz, chlorite, calcite, apatite, ilmenite, and sphene.

(b) Strongly altered dolerite

The more strongly altered dolerite occurs in the thinner dykes west of the Keith Metamorphics and most show evidence of shearing. The pyroxene is usually completely destroyed, and even much of the tremolite is chloritized. Fresh albite is uncommon, being completely altered to sericite. Scattered granules of zoisite are present. The doleritic texture is in places recognised, but is mostly destroyed by granulation and alteration.

The thin sheared dykes from Rocky Cape (33359, 33362, 33360, 33361) consist of a near-isotropic featureless mass of chlorite and sericite, with some quartz, ilmenite, sphene and leucoxene.

STRUCTURAL RELATIONS

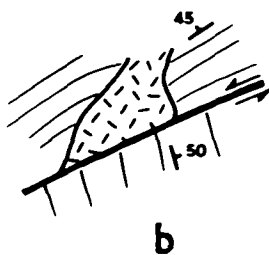
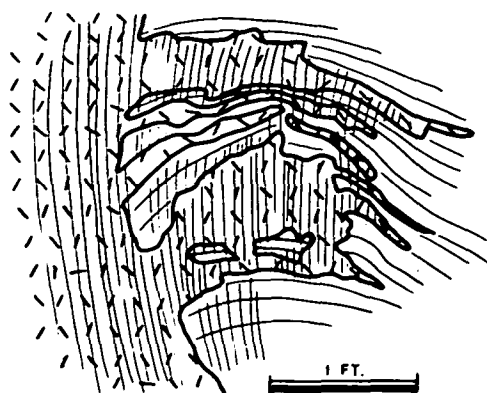
The dolerite bodies as a rule occur as dykes in the Rocky Cape Group (western assemblage), and as sills or slightly transgressive sheets in the Burnie Formation (eastern assemblage). Dolerite of this igneous phase has not been observed or reported in the Smithton Dolomite. However, it is expected that they should occur in the dolomite because the same structures that affect the dolomite also affect the underlying Rocky Cape Group. For example, dolerite is intruded along the axial plane of the syncline in the Cowrie Siltstone at Black River and the same syncline is found in the dolomite, Figure 17.

The petrological similarities suggest that all the dolerite bodies belong to the one petrogenic phase of intrusion, and the structural evidence suggests that the intrusions were essentially coeval or immediately post-dating the main (P1) phase of deformation.

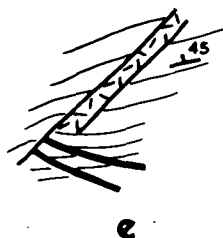
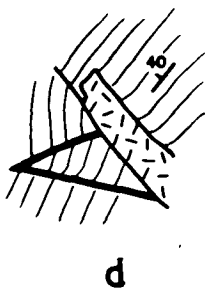
Dykes in the Rocky Cape Group

The dolerite bodies west of the Keith Metamorphics bear a clear and consistent relationship to the regional structure. These bodies are steeply dipping tabular dykes, the majority of which occur in the northeasterly trending high-angle thrust faults that are either located in, or parallel to, the axial planes of the folds. Some field sketches of dolerite outcrops on the shore-platform are shown in Figure 13. The location and trend of some of the larger dykes is shown on the structural map, (Figure 16). There are at least 30 smaller bodies up to 5 feet wide outcropping on the foreshore between Jacobs Boat Harbour and Black River. There are certain to be many more dykes than actually observed, especially inland, and it appears that they probably constitute a dyke swarm.

The thin dykes are sheared, and show clear evidence of having been intruded during the folding. In the small bay on the western side of Rocky Cape, opposite the old port is the Port Slate which is cut

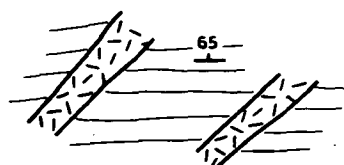
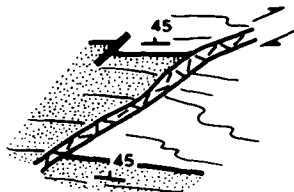


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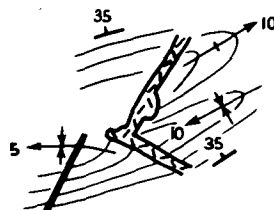
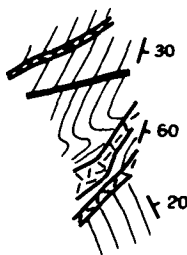
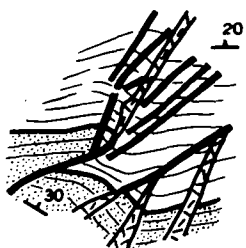
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0 200 400 600 ft.

Sketches of some dolerite dykes in the Rocky Cape Group.
 a. West side of Rocky Cape, b. Boat Harbour, c. East end of Sisters Beach, d. Wet Cave Point, e. Two miles west of Wet Cave Point, f, g, h. Port, Rocky Cape, i. West side of Rocky Cape, j. Edgcumbe Beach, k. Cowrie Point.

Figure 13.

by a prominent set of faults trending 080° . Intruded along one of these faults is a thin dyke 12 inches in width. A conspicuous shearing is parallel to the walls of the dyke. Within a stratigraphic interval of eight feet there are 12 sills in the enclosing flaggy quartzite and slate, emanating from the dyke. The sills are folded along with the bedding and possess a cleavage which is parallel both to the slaty cleavage in the siltstone, to the shearing in the dyke, and to the set of faults. The sills are mainly concordant with the bedding, but some smaller cross-cutting dykes occur which penetrate along the cleavage planes in the siltstone and along rotational joints in the sandstone, and small disoriented joint blocks are completely encircled by sheared dolerite. This is illustrated in Figure 13-a.

The shearing in the dolerite, which is congruent with the cleavage development in the enclosing rock, together with the penetration of dolerite along faults, cleavage and rotational joints which are all the result of folding, shows that the dolerite was intruded contemporaneously with folding.

The wider dolerite dykes up to 10 feet wide show a planar sheet jointing with slickensided faces, and a conspicuous planar shearing (33359) is visible in hand specimen. Such bodies possess a typical dolerite texture, but the feldspar and amphibole are almost

completely altered to sericite and chlorite. The pyroxene is commonly granulated. Coarsely spaced anastomosing fracture planes are visible in thin section, but are vague and indistinct due to a "healing" by growth of alteration products. It appears that alteration closely post-dated the shearing.

The larger dolerite bodies, between 10 and 300 feet wide, for example, at Detention River, are coarse-grained and massive, and show no evidence of deformation. It is improbable that such large bodies could have been intruded during folding and thrusting because of the high normal stress on the fault plane. These larger dolerite dykes were probably intruded along lines of crustal weakness after the relaxation of the local horizontal compression.

Sulphur Creek

Between Sulphur Creek headland and Howth is a sill about 40 feet thick which is cut and displaced by a set of faults that is genetically related to the first generation folds (Figure 57). The dolerite here was intruded before the first phase of folding.

Blythe Heads

Half a mile east of Blythe Heads is a small patch of sheared and altered dolerite. This is intruded into the plane of a fault which cuts and

displaces a fifth generation fold. This is the only structurally anomalous occurrence of dolerite. The P5 fold clearly refolds some P1 structures, so that this dolerite is younger than the P1 phase of folding. The exact age of the P5 structures is uncertain since it does not interact with any of the other fold generations, and is only tentatively assigned as the youngest in the fold sequence, (Table 6).

At a point 300 yards west of Blythe Heads are three sills, 20, 35 and 80 feet thick. These bodies are intruded along the bedding which is planar ^{generally} and vertical. There are no mesoscopic folds of any generation in the immediate vicinity and the dolerite is uncleaved so that the structural relations are not apparent.

Parklands

Dolerite outcrops extensively along a mile of foreshore between West Beach and Cooe Creek, immediately west of Burnie. This strip is termed the Parklands Flat Belt (Chapter 10) and contains abundant third generation folds and rare first generation folds. The slaty cleavage which normally pervades these rocks is not present in this belt. Figure 14 is a cross section of the dolerite bodies along the line shown on Figure 55. The inset in Figure 14 shows the suggested overall structure of the dolerite bodies. There appear

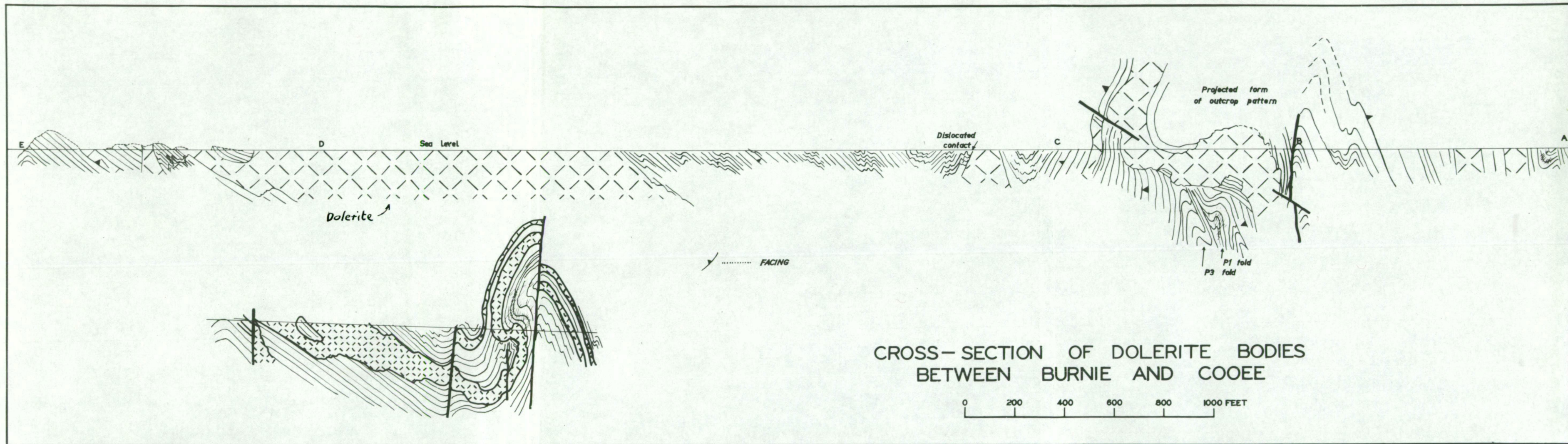


Figure 14.

? Legend.

to be two slightly transgressive, sill-like sheets which are folded by third generation folds.

✓ The largest mass at Parklands, is sill-like and is approximately 1000 feet thick. In detail, the lower contact is slightly transgressive and irregular with numerous offshoots in the form of sills, dykes and sinuous apophyses. Joints in these offshoots are continuous with the joints in the enclosing sedimentary rocks, and are probably related to the third generation of folding. These offshoots also behave independently of the joint planes in the host rock. This suggests that the intrusion was earlier than the third generation folding.

The arenite within 10 feet of the dolerite contacts assume a dull white glassy appearance in which the internal lamination becomes more conspicuous. In thin section the only change is a weak recrystallization of the matrix and the presence of limonite coating on the quartz grains. The individual quartz grains are unchanged.

About 50 feet above the base of the large body is a raft of sediment 100 feet long. In a zone five feet wide, inside the contact, the bedding laminations show extreme contortion with a "soft-sediment" style. The laminations are twisted and disrupted with no associated brecciation. Axial planes and fold axes are not consistently oriented, and many axial planes are

convoluted. There are three possible origins for this contortion. Firstly, it may be sedimentary contortion of the types described in Chapter 5. This is unlikely because it does not explain the close spatial relations. Secondly, it may be rheomorphic deformation associated with the intrusion. This is unlikely because of the unaltered nature of the contact rocks. Marked mineralogical change due to contact metamorphism would be expected. Thirdly, it may be due to intrusion while the sediments were relatively uncompact. A similar phenomenon has been noted by Morgan (1965), where quartz keratophyre intrudes unconsolidated sediment. This is the most likely explanation, and is supported by the sinuous apophyses at the basal contact of the main body.

Two sets of planar, acutely intersecting shear joints are present in the main body at Parklands. These joints are near-vertical, and trend northeast-southwest, parallel to the axial trend of the nearby P₃ folds. They intersect in a vertical axis. Slickensides with sub-horizontal plunges are present on the joint faces. This jointing has not been examined in detail, but it appears to be related to the P₃ folding, which also suggests the intrusion pre-dates the P₃ folding.

This suggestion is strengthened by the spatial relations between the dolerite and the third generation mesoscopic folds. The contacts are generally "welded" with no folding in the indurated zone, but in one locality (see Figure 14) where mesoscopic folds abut directly onto the dolerite, the contact is dislocated and there are no contact effects. The dolerite appears to have behaved rigidly without participating in the folding.

The dolerite body on the foreshore adjacent to the West Park Oval has an S-shaped outcrop pattern. In detail the contacts are irregular and suggest dilational openings with rupture across the bedding. When the outcrop pattern is projected back onto the plane of the cross-section, using the axis of the first generation fold (see Figure 14, and Figure 55), the form of the intrusion is concordant with the shape of the first generation coupled fold. A third generation fold (opposite vergence to P1, no cleavage and concentric style) is found beneath the P1 fold and in plan tapers toward the dolerite and vanishes at the contact. This again indicates that the dolerite bodies have acted as rigid massifs for the P3 folding.

In conclusion, the structural evidence indicates that the main period of intrusion of dolerite in the Burnie Formation is earlier than the P3 folding and

is related to the tectonic movements resulting in the main P1 phase of folding. It is not clear whether it is entirely pre-, syn- or post-tectonic. The dolerite bodies occur sporadically across the outcrop belt of the Burnie Formation, but are more common in the Parklands Flat Belt. This belt received virtually no deformation during the main P1 phase of deformation, but was only warped and possibly bodily translated. The main P1 phase of deformation is considered to be equivalent in age to the main deformation in the Rocky Cape Group. It therefore appears that the intrusions across the Rocky Cape Geanticline are almost all syntectonic to the main deformation of the Penguin Orogeny.

C H A P T E R 8

KEITH METAMORPHICS

The Keith Metamorphics, formally defined in Chapter 2, is a belt of low-grade regional metamorphic schist and amphibolite which separates the Rocky Cape Group to the west and the Burnie Formation to the east. The schists are derived in part from the surrounding sediments, and in part from the syntectonic basic intrusions considered equivalent to the Cooe Dolerite. The belt has been followed from Wynyard on the coast, southwest to the Arthur River and probably extends much further.

Outcrop generally is sparse and deeply weathered. Only in the bed of the Inglis and Flowerdale Rivers is fresh exposure found.

MINERALOGY

The Keith Metamorphics contain quartz, albite, calcite, muscovite, chlorite, tremolite and epidote as the essential minerals, and less than 10% of biotite, zoisite, sphene, magnetite, ilmenite, leucoxene, tourmaline, zircon and apatite.

Plagioclase occurs in large (up to 2mm) porphyroblastic crystals or as shapeless grains in a granoblastic aggregate. It ^{has} is abundantly ^{sons} included with fine amphibole, epidote, zoisite, mica and zircon, which are commonly arranged in ill-defined sigmoidal curves.

? possibly different
metamorphic facies
considered separately?

The plagioclase is nearly the pure albite end-member. Eight determinations from specimens 33237 and 33239, using the method of Slemmons (1962), gave results in the narrow range An 0% - 3%. It has a relief less than balsam and 2V of 82° - 85° . It generally shows simple twinning according to the Albite-Law and rare polysynthetic twinning.

Actinolite occurs in small fibrous needles and laths. It has a maximum $Z^{\wedge}c$ of 20° . It is generally pale green but in some specimens (33229, 33241) has a faint blue-green pleochroism which may indicate a tendency toward a sodic-rich amphibole.

The micaceous minerals which define the foliation consist of muscovite, chlorite and biotite. The chlorite occurs in well-formed, slightly pleochroic green flakes and is a primary metamorphic mineral rather than a hydrothermal or retrogressive alteration product. It is uniaxial positive with a medium relief, very low birefringence and has the properties of prochlorite. The biotite occurs in minor quantities as small ragged pleochroic flakes or as birefringent bands within the chlorite (33234).

Epidote occurs in abundant small disseminated granules, (33242) or more rarely in aggregates of large crystals (33241).

The more basic schist is characterized by the abundance (locally up to 10%) of sphene, magnetite and ilmenite. Magnetite occurs in crystals of square, triangular or rhombic section, often deeply embayed with sphene and quartz. Ilmenite forms large skeletal blades up to 3mm in length.

PETROLOGY

Three main rock assemblages occur in the Keith Metamorphics.

- (a) pelitic and calcic schists
- (b) albite amphibolite
- (c) transitional rock-types

(a) Pelitic and calcic schists

Medium to coarse-grained, quartz-muscovite - (albite-calcite) schist is common. Specimen 33249 (Plate 18-a) from the Arthur River is typical of the finer grained foliated schist. It consists of muscovite and quartz with an ill-defined segregation into layers up to 0.3mm across. The quartz grains average 0.02mm, and are interlocking. Minor amounts of cross-fibre porphyroblastic chlorite (penninite?) and faintly pink pleochroic zoisite are wrapped by the foliation. A chemical analysis is given in Table 5.

Specimen 33235 from the Inglis River is a crenulated muscovite schist. It is composed of strongly oriented muscovite and flattened quartz grains which

define the strong foliation. There is little segregation in the foliation plane. The crenulation is the expression of the strain-slip cleavage which is common in these rocks.

Specimen 33240 from the Flowerdale River is a green and white striped, well-foliated schist with a later crenulation. The lighter coloured layers are rich in quartz and the darker layers are rich in muscovite plus epidote, sphene, chlorite, tourmaline and magnetite. Albite porphyroblasts up to 0.5mm are common and are confined to the muscovite-rich layers. The muscovite does not show a high degree of preferred orientation within the foliation. Muscovite crystallization may therefore be post-kinematic. Specimen 33233 from the Inglis River is similar, but contains about 10% of albite up to 1.5mm, and 10% calcite.

Specimen 33239 from the Flowerdale River is a well-foliated quartz-muscovite - calcite-albite schist. It has a segregation into quartz - calcite rich layers, and chlorite-muscovite rich layers which contain about 5% of albite. The calcite is sub-idioblastic and varies in size up to 2mm. Albite, up to 1.5mm, has a porphyroblastic habit, being wrapped by the foliation. Trails of quartz inclusions in the albite are straight or gently sigmoidal and are continuous with the mica foliation. A chemical analysis of 33239, shown in

Table 5

ANALYSES OF KEITH METAMORPHICS

	1	2	3	4
SiO ₂	73.4	55.0	46.2	57.6
Al ₂ O ₃	11.2	12.0	10.8	10.2
Fe ₂ O ₃	4.0	2.1	10.1	4.4
FeO	2.0	5.1	7.7	4.2
MgO	2.0	5.7	6.0	4.8
CaO	0.20	5.6	8.4	4.8
Na ₂ O	0.07	1.1	3.4	3.9
K ₂ O	3.4	2.8	0.60	1.2
H ₂ O-	0.15	0.20	0.17	0.14
H ₂ O+	2.5	3.0	2.6	1.5
MnO	0.01	0.12	0.31	0.06
TiO ₂	0.46	0.36	3.0	0.61
P ₂ O ₅	0.06	0.07	0.22	0.32
CO ₂	-	7.5	-	5.8
SO ₃	0.08	0.16	0.21	0.09
S	0.03	-	0.19	-
	99.56	100.81	99.90	99.62
	-----	-----	-----	-----

1. Fine-grained schist, Hilder's timber road, Arthur River, 33247.
2. Quartz-muscovite-calcite-albite schist, Flowerdale River, 33239.
3. Albite amphibolite, Flowerdale River, 33241.
4. Albite amphibolite, Flowerdale River, 33237.

Analyst : H.K. Wellington

Table 5, shows a marked excess of K_2O over Na_2O .

(b) Basic schist and albite amphibolite

These rocks vary from well foliated greenschists, to weakly foliated, massive, dense, dark-green rocks. They are characterized by the abundance of albite, the presence of actinolite which in places shows a blue+green pleochroism, and high Na_2O content.

Specimen 33229 from the Inglis River is a mottled greenschist with abundant black idioblastic magnetite and ilmenite visible in hand specimen. It is composed dominantly of equal proportions of albite, quartz, epidote and tremolite. Albite and quartz are up to 0.75mm, and form a xenoblastic mosaic. The actinolite in places shows a blue-green pleochroism. Epidote, amphibole, and sphene commonly occur in shredded aggregates which define the foliation. The magnetite has a porphyroblastic habit. Chlorite forms about 5%, and tourmaline and rutile are accessories.

Specimen 33241 from the Flowerdale River contains 50% of albite, tremolite 20%, epidote 15% and quartz. Albite and quartz occur in equant xenoblastic grains up to 0.5mm, forming a mosaic which is abundantly studded with actinolite and granular epidote. The inclusions in the albite show a regular, sigmoidal arrangement. The foliation is defined by strewn-out aggregates of sphene and leucoxene, shreds

of granular epidote (0.02mm), and dimensionally oriented sheaths of amphibole. Blue-green pleochroic actinolite is again present. A chemical analysis of this specimen given in Table 5 shows a low SiO_2 content, and a very high Na_2O content.

Specimen 33242 from the Flowerdale River is a finer grained variety composed of a granular mosaic of quartz and albite (0.1 - 0.05mm), and characterized by small randomly oriented idiomorphic prisms of amphibole up to 0.5mm in length. The amphibole occupies about 15% of the rock. Small granules of epidote are common. Tourmaline and sphene are accessories.

Some of the more dense, massive rocks are very rich in albite. Specimen 33237 from the Flowerdale River is a dark gray-green dense rock with a vague indistinct foliation due to lighter coloured albite-rich layers and darker chlorite-rich layers. It is composed of about 70% albite in equidimensional grains (0.5 - 1.0mm) forming a xenoblastic mosaic, (Plate 18b). The albite has abundant inclusions of quartz, calcite, chlorite, muscovite and magnetite, arranged in straight or slightly curved trails. Calcite, quartz and green biotite are the other main constituents. Magnetite and ilmenite together form about 5% of the rock. The chemical analysis of Specimen 33237 (Table 5) shows it to be very high in Na_2O content. Specimen 33238 from the Flowerdale River

is similar, and contains abundant ilmenite, magnetite, hematite and pyrite.

Of particular importance are some of these albite-rich amphibolites which show remnant dolerite textures . These are found in the Inglis River, one mile downstream from the Caulder-Preolenna Road bridge. These are fine-grained or aphanitic, dense green rocks, with diffuse pink albite spots visible in the hand specimen. Apart from joints and thin pink albite veins, they are massive and devoid of foliation. The relationships with the surrounding pelitic schist and greenschist is uncertain, but they appear to be near-vertical tabular bodies concordant with the foliation.

In thin section (33231, 33234) these rocks are heavily altered and their original mineralogy is difficult to deduce. They consist of quartz, plagioclase, actinolite, magnetite, apatite, sphene, epidote and zoisite. The veins cutting the rock are composed of euhedral albite, with inters_titial quartz, calcite and magnetite, and are thus similar to the leucocratic veins in the massive dolerite intrusions at Cocee.

The groundmass of specimen 33231 contains about 40% albite in randomly oriented interlocking subhedral laths up to 0.4mm. The subhedral albite has polysynthetic twinning on the Albite Law. Albite also occurs in small

shapeless intergranular crystals forming a mosaic with the quartz. These smaller albite grains are abundantly included with zoisite, epidote, tremolite, calcite and sphene. The remnants of the ferro-magnesian minerals are now in the form of large columns (0.5mm) of interlayered fibrous calcite, amphibole and chlorite (penninite?). These may be pseudomorphs of original pyroxene. Small flakes of chlorite (prochlorite?) are interstitial to the albite laths. Apatite needles up to 0.5mm long, skeletal ilmenite, and sphene are the accessories.

Specimen 33234 consists entirely of material similar to the finer grained patches of albite mosaic in the previous specimen (33231). It consists of about 60% small albite (0.01mm) having vague and indistinct borders, with some interlocking quartz. Albite is included with calcite, biotite and other minute semi-opaque minerals. Abundant flakes of small, dirty green biotite, chlorite, irregular patches of calcite, and minor granules of zoisite are interstitial to the albite. Long needles of apatite (0.8mm) and skeletal ilmenite are common accessories.

The rocks showing these remnant doleritic textures, are similar to parts of the altered but unsheared dolerite from Coocoo. Commonly the dolerite shows a patchy development of a mosaic texture of interlocking, shapeless, equant albite grains.

(c) Transitional rock-types

The western contact between the coarsely crystalline schist and the unmetamorphosed orthoquartzite of the Rocky Cape Group to the northwest shows gradational properties across the strike. Rock types in this zone include well-foliated phyllite, schistose quartzite with mortar texture, and slate interbedded with cleaved arenite. For the purpose of field mapping the transitional rock types are divided from the crystalline schist by the first appearance of albite in hand specimen. This also corresponds to the first appearance of albite on the microscopic scale. The "albite-line" trends northeast-southwest, parallel to the regional trend of the belt of metamorphic rocks and parallel to the trend of the foliation within the metamorphic rocks.

Specimens 33226, 33227 are typical of the phyllite. They consist mainly of muscovite (50%), quartz (40%), with minor amounts of chlorite (penninite?), tourmaline, zircon and sphene. The foliation is defined by a high degree of preferred orientation of small crystalline muscovite, a distinct dimensional orientation of small flattened quartz, and also in part by a segregation into muscovite and quartz-rich layers. The flattened quartz grains are elongate by as much as 5:1 in the plane of the foliation and average 0.3mm in smallest diameter. Away from the later kinks, these flattened quartz grains show

no undulose extinction and muscovite penetrates the quartz in the pressure shadows, indicating that recrystallization was an important factor in the formation of the foliation.

Interlayered in the marginal phyllite in the western transition zone are two horizons of sheared quartzite which show cataclastic mortar texture. These are about 100 feet thick and are considered to represent original beds.

The quartzite horizon closest to the crystalline metamorphics is just outside (to the west) the "albite-line" but is still one mile from the nearest unmetamorphosed orthoquartzite of the Rocky Cape Group. The marginal quartzite is a pale green, fine-grained vitreous rock (33243, 33244), with a bedding foliation and some intrafolial isoclinal folds. Mortar texture has begun to develop but appears to have been arrested by slight syntectonic crystallization. The larger parent grains (0.15mm) show undulose extinction, fracturing, deformation bands and border granulation. The boundaries are minutely sutured. The ground mass is composed of a fine, closely interlocking mosaic of granular (0.01mm) strain-free quartz, indicating some syn- or post-tectonic recrystallization. In thin section there is a weak foliation due to dimensional orientation of small flakes of chlorite, muscovite and

biotite, and also by shreds of mica which anastomose and enclose clusters of larger parent grains.

Specimens 33224 and 33225 are from the quartzite horizon further to the west, that is, closer to the unmetamorphosed rocks. These possess a strong foliation parallel to the bedding which is the counterpart of the foliation in the phyllite. In hand specimen, they are strongly sheared, brown-coloured rocks, superficially resembling finely knotted albite schist, but in thin section these knots are seen to be large quartz porphyroclasts which range from 0.1 mm to 1.3mm. The foliation is due to discontinuous anastomosing shreds of fine muscovite and minor biotite which enclose lozenge-shaped slices containing one or more quartz porphyroclasts. Coarser crystalline muscovite and small biotite have grown in the pressure shadows and penetrate the quartz. The porphyroclasts are slightly elongate (about 3:1) in the foliation, but some are rotated out of this plane producing oblique curved pressure shadows.

Some of the larger porphyroclasts have a zonal arrangement, consisting of a large, strained quartz grain with a sutured margin, a fringe of finely granulated quartz, then a groundmass of quartz, and micaceous material. (Plate 19-a). This suggests that the granular quartz is due in part to border granulation by shearing, and in part to original recrystallized

sedimentary matrix.

The anastomosing cleavage may not entirely be due to the influence of individual grains because in some cases, several clastic grains are enclosed within the cleavage slices. In specimens 33224 and 33225, the anastomosing nature is so regular that there appears to be two acutely intersecting cleavage sets. This may be the more intense expression of a rhombic cleavage pattern in the arenite on the western margin of the transition zone, south of Lapoinya. This rhombic cleavage is due to the interaction of a bedding fissility and a sandstone cleavage oblique to bedding. With increased flattening by granulation and recrystallization, a penetrative anastomosing cleavage inclined at only a small angle to bedding, may be formed.

METAMORPHISM

Metamorphic facies

The mineral assemblages in the basic schist, (albite-epidote-chlorite-actinolite-sphene-calcite, and less commonly albite-epidote-biotite), indicate that the basic schist has reached the lower and middle subfacies of the greenschist facies of regional metamorphism. The soda-rich plagioclase ($An_{0\%}-An_{3\%}$), together with the absence of hornblende amphibole and almandine garnet show that the metamorphism has gone no higher. This is also compatible with the mineral assemblage of quartz-muscovite-chlorite (-biotite-calcite) in the pelitic and calcic schists.

The mineralogy of these rocks forms a compatible assemblage with no evidence of an earlier phase of metamorphism. Similarly, the textures indicate a simple metamorphic history. The main mesoscopic foliation is due to a segregation into layers of a quartz-calcite-albite mosaic, and to layers of chlorite and muscovite with a preferred orientation. This foliation is the only one visible microscopically, there being no remnants of an earlier tectonic foliation. Albite with regular trails of inclusions is wrapped by the foliation. These trails are straight, gently curved or slightly sigmoidal and are congruent with the foliation.

These features indicate that there was only one period of metamorphism and penetrative deformation which affect^{ed} the Keith Metamorphics. This contrasts with the polyphase history of the older Precambrian basement in the Frenchmans Cap area described by Gee (1963) and Spry (1963).

Origin of Amphibolite

In considering the origin of the basic schist and amphibolite there are two possibilities :

Dolomitic mudstone : Some of the mineral assemblages are capable of being derived from an impure dolomite, and dolomitic mudstone occurs in the Irby Siltstone at Sisters Beach. The structural and stratigraphic correlations (Figure 16) are such that this horizon would pass well to the west,

of the Keith Metamorphics, on the western side of the prominent line of orthoquartzite hills (Jacob Quartzite) that extends from Jacobs Boat Harbour to the Arthur River. Furthermore, this interpretation does not explain the abundance of ilmenite, sphene, magnetite and albite. The chemical analyses in Table 5 contrast with that for the dolomite at Sisters Beach, (Page 56). Even allowing for an impure dolomite, this interpretation does not explain the relatively high proportions of FeO , Fe_2O_3 , TiO_2 , and Na_2O .

Intermediate or basic igneous : Several factors suggest that the basic schist and amphibolite were derived by metamorphism of igneous rocks. These include the dense massive appearance in outcrop, the high proportion of ilmenite, magnetite and sphene, and the remnant doleritic textures with associated needles of apatite in the unfoliated rocks. The most obvious source rock is the albite dolerite (Cooee Dolerite) which was intruded into both the eastern and western sedimentary assemblages. These are believed to have ^{been} intruded syn-tectonically to the main phase of deformation. Minor quantities of pillow lavas are known from the Burnie Formation, but are not known to occur in sufficiently large masses.

It is therefore postulated that the schists are derived by regional metamorphism of the Coöee Dolerite and the enclosing sedimentary material. There is a close similarity in texture of the more strongly altered dolerite at Coöee and the more massive amphibolite from the Inglis River. They both possess a mosaic of equant shapeless albite grains, heavily included with the products of retrograde metamorphism or saussuritization. There is also a close similarity in chemical composition between the Coöee Dolerite in Table 4 and the albite amphibolite in Table 5. This is reflected in the similar mineralogy.

Transitional nature of the western contact

The Keith Metamorphics is gradational across strike into the unmetamorphosed Rocky Cape Group to the west. This is expressed in two ways.

1. There is a gradation in metamorphic grade from orthoquartzite showing no recrystallization and poorly cleaved siltstone through slate, phyllite, medium-grained albite-chlorite schist, into coarse-grained schist with biotite.
2. There is a gradation in fabric development from textures in the orthoquartzite showing no penetrative deformation on the granular scale; through cleaved slightly granulated arenite; through mortar-texture quartzite showing penetrative granulation,

and then into rocks showing dominantly recrystallization textures. This is reflected in the mesoscopic aspect of the foliation. There is no cleavage in the Jacob Quartzite. In the interlayered arenite and slate lying stratigraphically above the Jacob Quartzite, the cleavage is similar to that in the Burnie Formation, being a slaty and sandstone cleavage oblique to bedding. Sliding along bedding and oblique cleavage is common. Closer to the metamorphic rocks, the arenite acquires an anastomosing cleavage producing lozenge-shaped slices, slightly acute to bedding. This appears to be due to shearing and growth of mica along bedding and oblique cleavage. The finer grained beds become lustrous phyllite with a barely visible segregation. In the vicinity of the "albite-line", the phyllite becomes coarser, so that individual mica flakes are just visible with the naked eye, and albite spots appear. To the east of the "albite-line", the schist is coarser, with readily visible albite porphyroblasts up to 1.5mm.

This gradation is seen in the forestry road (Hebe Road) south of Lapoinya, near the junction of the Eiby and Flowerdale Rivers. In this section, the oblique cleavage in the transitional beds is lying more shallow than bedding. The transition from virtually unsheared quartzite, through sheared arenite and slate, into phyllite and fine-grained schist is seen on Milders Road between Meunna and the Arthur River.

In this section, ripple marks are preserved in the arenite interbedded with slate. The passage across the "albite-line" is best seen in the Flowerdale River, two miles upstream from the Pages Road bridge.

The eastern contact with the Burnie Formation is covered by Permian and Tertiary rocks. However, it suggested that it is also gradational across strike because there is a gradual attenuation in intensity of the main cleavage in the Burnie Formation, eastward away from metamorphic belt.

Although there is a clear gradation on the western contact of the Keith Metamorphics, this does not prove conclusively that the metamorphics were derived from the adjacent sedimentary rocks. To do this it is necessary to demonstrate a gradation along the strike. The alternative is that the gradation is due to a series of steps, each step marking a thrust which brings up from depth, slices of older Precambrian basement. Such a configuration is untenable on structural grounds as explained later in this chapter.

STRUCTURE

Mesoscopic structure

This belt of metamorphic rocks is dominated by a strong near-vertical regional schistosity, trending parallel to the belt. In the pelitic schist

and phyllite there is also an earlier diffuse banding which is the folded surface. This layering, seen only on the mesoscopic scale, is coarser than the segregation along the foliation, and is considered to be bedding. A lineation is produced by the intersection of these two surfaces. Occasional folds in the layering have axes parallel to this lineation. No such layering is seen in the greenschist. Orientation data for the foliation, lineation, and earlier layering are shown in Figure 15-a.

Virtually nothing is known of the internal structure of the metamorphic belt. The amphibolite appears to form steeply dipping tabular bodies, concordant to the northeast-southwest trending foliation. This is analogous to the dominant structural configuration of the dolerite bodies in the Rocky Cape Group which are northeast-southwest trending, steeply dipping, tabular bodies intruded along axial-plane break thrusts and cleavage planes, and are commonly sheared.

Late strain-slip cleavage

A late strain-slip cleavage is well developed in the phyllite in the western transition zone, and poorly developed in the coarser pelitic schist. This cleavage is generally near-vertical, and thus intersects the earlier foliation at small to moderate angles. In the hand specimen it appears as regularly spaced (0.5 - 2.0mm apart) planar slip planes.

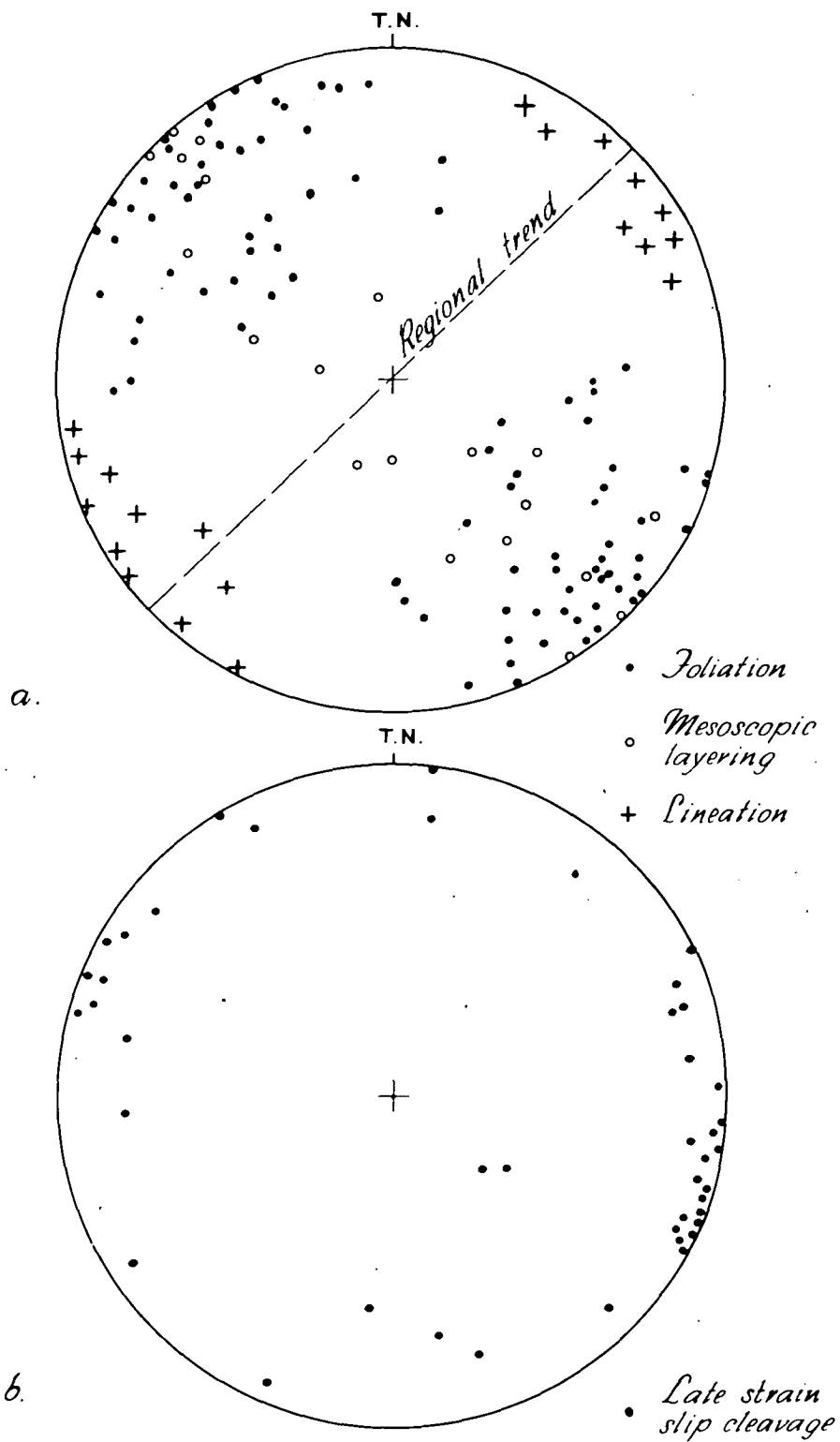


Figure 15.

In thin section the slip surfaces consist of many discontinuous kink planes, often acutely intersecting or arranged en échelon. The slip surfaces are not exactly discrete planes of movement, but rather domains of movement in which the micas of the pre-existing foliation have been mechanically rotated almost into the plane of movement (Plate 19-b). Thus it is the micro-crenulations that produce the planes of slip and not vice versa. There is no mineral growth along the new cleavage planes, and there has been no associated recrystallization, because lenticular quartz, mica and oriented tourmaline are bent and broken. The strain-slip cleavage generally maintains a constant sense of displacement in the one specimen, although in thin section minor bends with the opposite displacement are found. These antithetic slips are inclined at up to 20° to the mean cleavage plane.

The plot of poles to strain-slip cleavage (Figure 15-b), taken mainly from the phyllite of the western transition zone, indicates a mean trend of 015° which is oblique to the main foliation. It appears to be related to concentric style undulations of the earlier foliation.

This strain-slip cleavage is similar in style and orientation to the P^4 cleavage found in the

more lustrous slate in the Burnie Formation between Doctors Rocks and Cam River. This is a later cleavage associated with concentric undulations of the bedding, (Table 6).

Major Structure

The Keith Metamorphics is a belt of sheared and metamorphosed sedimentary rock and basic intrusives which separate two different suites of deformed Proterozoic sediments. Its exact width is not known, because the eastern contact is obscured by Permian sediments, but it has a minimum width of $3\frac{1}{2}$ miles and a maximum possible width of 7 miles.

Over the extent of the western contact, from Wynyard to the Arthur River the foliation in the metamorphics is parallel to the bedding and to the regional fold axis in the sediments to the west. The long strike ridge of unmetamorphosed orthoquartzite (correlate of Jacob Quartzite) that flanks the entire length of the western contact is mainly vertical or overturned and faces downward to the southeast toward the Keith Metamorphics. Conclusive facings can readily be obtained from the orthoquartzite. Good examples of overturned cross-bedding can be found one mile southwest of the Forestry Commission house at Lapoinya.

This important facing means that the Rocky Cape Group stratigraphically underlies the Keith Metamorphics although it partly overlies the Keith Metamorphics in a structural sense. It follows that the whole of the Rocky Cape Group succession must turn down at the western contact of the Keith Metamorphics. Even allowing for a substantial thinning of the Rocky Cape Group succession, it is not possible to bring it back up and over the Keith Metamorphics. The distribution of rock types, and the persistence of the down-to-the-east facing throughout the transition zone (as shown by the relationship of bedding and cleavage) show that there is no such reversal.

On the eastern side, the Burnie Formation is overturned and faces downward to the southeast. Thus the Keith Metamorphics stratigraphically underlies the Burnie Formation but actually overlies it in a tectonic sense.

This configuration, illustrated in the composite profiles of Figure 52 and Figure 53, rules out the possibility that the metamorphic belt is an older basement high, against which the Rocky Cape Group sediments were deposited. The simple asymmetrical arrangement, involving overfolding and high-angle thrusting to the southeast also rules out the possibility of bringing up from depth imbricate

thrust slices of older basement during crustal deformation, since the sense of movement on these thrusts would be the opposite to that demanded by the overall couplet.

All the structural evidence points to the Keith Metamorphics being a dispersed zone of high-angle thrusting, formed as the unstable-shelf facies to the west was folded and transported to the southeast, squashing the flysch-type sediments against the Precambrian metamorphic basement which lies at least 40 miles to the east. The metamorphic belt appears to mark a line of deep crustal weakness which acted as a conduit for the intrusion of dolerite. It was thus probably a "hot-spot", as well as a line of intense shearing. It therefore probably marks a zone of incipient deep-seated regional metamorphism which usually pervades the entire sedimentary pile in larger geosynclines.

THE ARTHUR LINEAMENT

These metamorphic rocks can be correlated with other rocks outside the area studied. They can be correlated directly, by walking, with the Keith Beds of McNeil (1961), from which the name Keith Metamorphics is defined (Chapter 2).

Spry (1964) has described the petrology of the Whyte Schist, a belt of metamorphosed sediments and igneous rocks about five miles wide, in the lower

Pieman River Area. This belt trends approximately north-northeast from Granville Harbour and is flanked to the east and west by large thicknesses of unmetamorphosed Precambrian sediments. The Whyte Schist shows a remarkable degree of similarity, in petrology, chemical composition, and metamorphic grade, to the Keith Metamorphics. Several dykes of altered igneous rocks occur in the pelitic schist in the lower Pieman River area. These contain a mineral assemblage, identical in every respect with the amphibolite in the Flowerdale and Inglis Rivers.

The Whyte Schist extends northward into the Savage River area (Urquhart, 1966), where it is a meridional belt of pelitic and psammitic schist with amphibolite bodies and magnetite lenses. The mineral assemblages and metamorphic facies are also identical with that of the Keith Metamorphics. Urquhart (1966) also noted the cataclastic textures.

The intervening twenty miles between the Savage River iron ore deposits and the Arthur River is geologically unknown, but because of the large scale of the structure, the correlation of the Keith Metamorphics with the Whyte Schist appears justified. This is strongly supported by the aeromagnetic survey of west and northwest Tasmania conducted by RioTinto Australian Exploration (1956) which shows a continuous line of anomalies along the suggested line. This lineament therefore appears to be a ribbon-like belt

of sheared metamorphic rocks stretching for 60 miles in more or less a straight line from Granville Harbour to Wynyard across the northwest corner of Tasmania (Figure 51).

It is proposed to call this line the Arthur Lineament and it appears to represent the fundamental tectonic structure of the Upper Proterozoic of Tasmania.

The correlation of the Whyte Schist and the Keith Metamorphics has an important bearing on some structural and stratigraphic interpretations in the Precambrian rocks of northwest and west Tasmania, and also raises some points of conflict with some previous ideas.

Spry (1964) considered the Whyte Schist to be a remnant basement high of older Precambrian rock against which the unmetamorphosed sediments were deposited. The metamorphism and deformation was taken to record the older Frenchman Orogeny, a period of widespread regional metamorphism affecting the older Precambrian basement in the Tyennan Nucleus. This interpretation was based mainly on microtextural evidence of the pelitic schist which contain remnants of a tectonic surface earlier than the main foliation. Similar textures are found in the older Precambrian basement in the Frenchmans Cap area (Spry 1963, Gee 1963), but have not been found in the Keith Metamorphics. However their existence is not sufficient

ground for correlation since polyphase movements may well occur within the rocks of the Arthur Lineament. It has also been suggested in this chapter that the dominant foliation in the Keith Metamorphics is due to flattening of lenticular slices defined by a bedding cleavage and the oblique cleavage. It is possible that some of the earlier surfaces, recorded by Spry (1964), may be such a feature.

Spry (1964, p.44) discussed the possible equivalence of the amphibolite in the Whyte Schist and the sodic dolerite in the unmetamorphosed sediments to the west, but preferred to assign the amphibolite to the Frenchman Orogeny and the dolerite to a later age. None of the field evidence presented by Spry rules out this correlation. Spry also considered that the dolerite was distinct in chemical composition from the amphibolite, especially on Na_2O content. It has been pointed out in this chapter that the two rock types are chemically similar, and the Na_2O contents are not significantly different. There is, however, a marked contrast in the Na_2O and K_2O contents between the amphibolite of the Whyte Schist and the amphibolite found elsewhere in the older Precambrian metamorphic basement (see tables in Spry, 1962^a, p.281).

In support of the extension of the ideas put forward in this thesis, Spry (1964) recorded that :

- (a) The eastern contact of the Whyte Schist is apparently transitional into the Oonah Quartzite (p.44),
- (b) the trend of the dyke swarm to the west is parallel to the trend of the Whyte Schist belt, (Figure 1),
- (c) the unmetamorphosed rocks to the west are dipping toward, and under, the Whyte Schist, (p.30, figure 4); and those to the east are dipping away, and off, the Whyte Schist, (p.44).

This suggests that a complete re-appraisal of the stratigraphic succession in the Proterozoic in the Lower Pieman River area is necessary. This is discussed in the final chapter.

In the northerly extension of the Whyte Schist in the Savage River area, Urquhart (1966, p.38-39) pointed out the chemical contrast between the Savage River amphibolite and the amphibolite of the older Precambrian metamorphic basement. However, he further considered that the Savage River amphibolite was chemically different from the dolerite. Urquhart (1966) then correlated the Savage River amphibolite with the Cambrian spilitic rocks on purely chemical grounds, despite the fact that the amphibolite has the same metamorphic grade as the enclosing metasediments which he considered to record a metamorphic period of Precambrian age. However the degree of

alteration in the dolerite and the amphibolite makes correlations on chemical grounds uncertain and is unsupported by the present work.

There are several recordings of a probable gradational boundary between this same metamorphic belt and the surrounding sediments. For example, McNeil (1961, p.50) notes a gradation between the Neasy Quartzite and the metamorphosed Keith Beds. Blissett (1962, p.21) at Zeehan suggested a gradational contact between the Oonah Quartzite and the Whyte Schist north of Duck Creek in Zeehan Quadrangle. Similarly, Spry (1964 p.44) stated that the contact between the Whyte Schist and the Oonah, which is exposed in the Pieman River, appears to be gradational over half a mile.

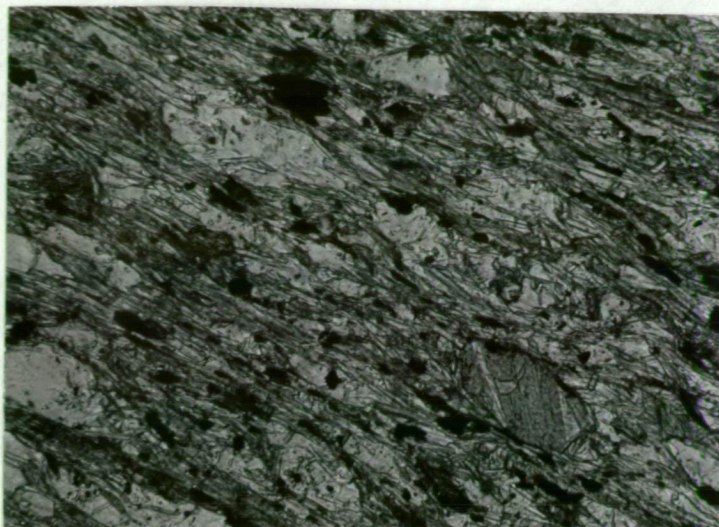


Plate 18a Photomicrograph of fine-grained
quartz-muscovite schist with
porphyroblastic chlorite, Keith
Metamorphics, Arthur River.
Sp. 33249, x 150.

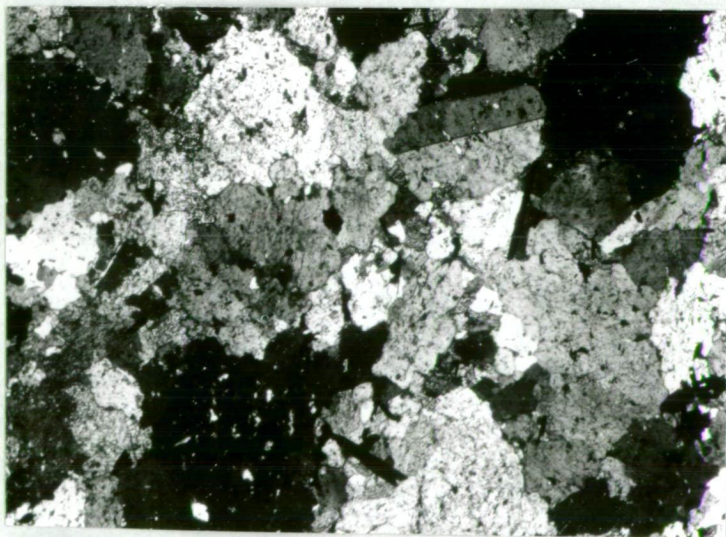


Plate 18b Photomicrograph of albite mosaic
in albite-rich schist, Keith
Metamorphics, Flowerdale River.
Sp. 33338, x 30.

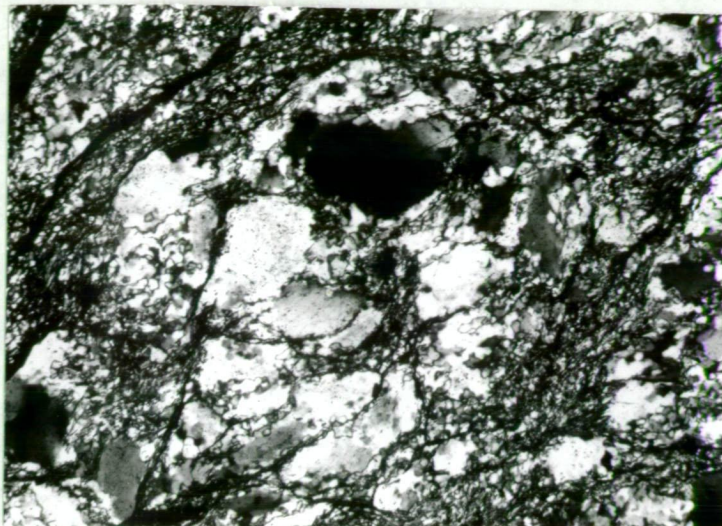


Plate 19a Photomicrograph of mortar texture
in quartzite, showing zonal
arrangement of porphyroclasts,
western margin of Keith Meta-
Morphics, Flowerdale River.
Sp. 33225, x 1000.



Plate 19b Photomicrograph of strain-slip
cleavage in phyllite in western
margin of Keith Metamorphics.
Sp. 33235, x 23.

C H A P T E R 9

STRUCTURE OF THE ROCKY CAPE GROUPREGIONAL STRUCTURE

The main structural feature of that part of the Rocky Cape Geanticline to the west of the Keith Metamorphics is a series of northeast-southwest trending anticlines and synclines which are tight and asymmetrical near the Keith Metamorphics, and become broad and more symmetrical further to the west. These structures are cut and displaced by a major transcurrent fault with a dextral displacement of about five miles.

The lithologic trends and major structural elements are shown in the structural map (Figure 16). Some of the more important folds have been named in order to facilitate description of the reconstruction before the transcurrent faulting. Adjacent to the Keith Metamorphics the regional strike of bedding is northeast-southwest because of the tight folding. The correlate of the Jacob Quartzite can be followed in a straight line from Jacobs Boat Harbour for twenty miles to the southwest over the Arthur River to Folly Hill. It dips steeply to the northwest, and cross-bedding shows that it is overturned, and faces downward to the southeast. The Dip Range anticline, which flanks the western boundary of the Keith Metamorphics, is therefore asymmetrical and overturned

against the Keith Metamorphics. The next anticline to the west is the Sisters Hills anticline which has an axial plane dipping 70° northwest but has no associated overturning. The next fold to the west is the Newhaven syncline, which is a more open fold, and the lithological trends swing to a northwest-southeast direction. Further to the west is a series of second-order broad open folds with gentle northeasterly plunges. These folds persist to the northwest at least as far as Black River, where they affect the conglomerate and dolomite (Figure 17). The folding in the Rocky Cape Group therefore appears to be later than the deposition of the Smithton Dolomite.

The overall structure west of the Keith Metamorphics is shown in the left-hand half of Figure 52. This section is a composite of axial projection profiles, and projected cross sections along the tectonic axis. The effect of cross faulting has been removed and the structure freely sketched in where superficial rocks cover the Precambrian. The structure is somewhat diagrammatic but all major features can be substantiated.

The style of the major folds reflects the lithology of thick massive orthoquartzite with thinner siltstone layers. The folds are characterized by nearly planar limbs and axial-plane break thrusts.

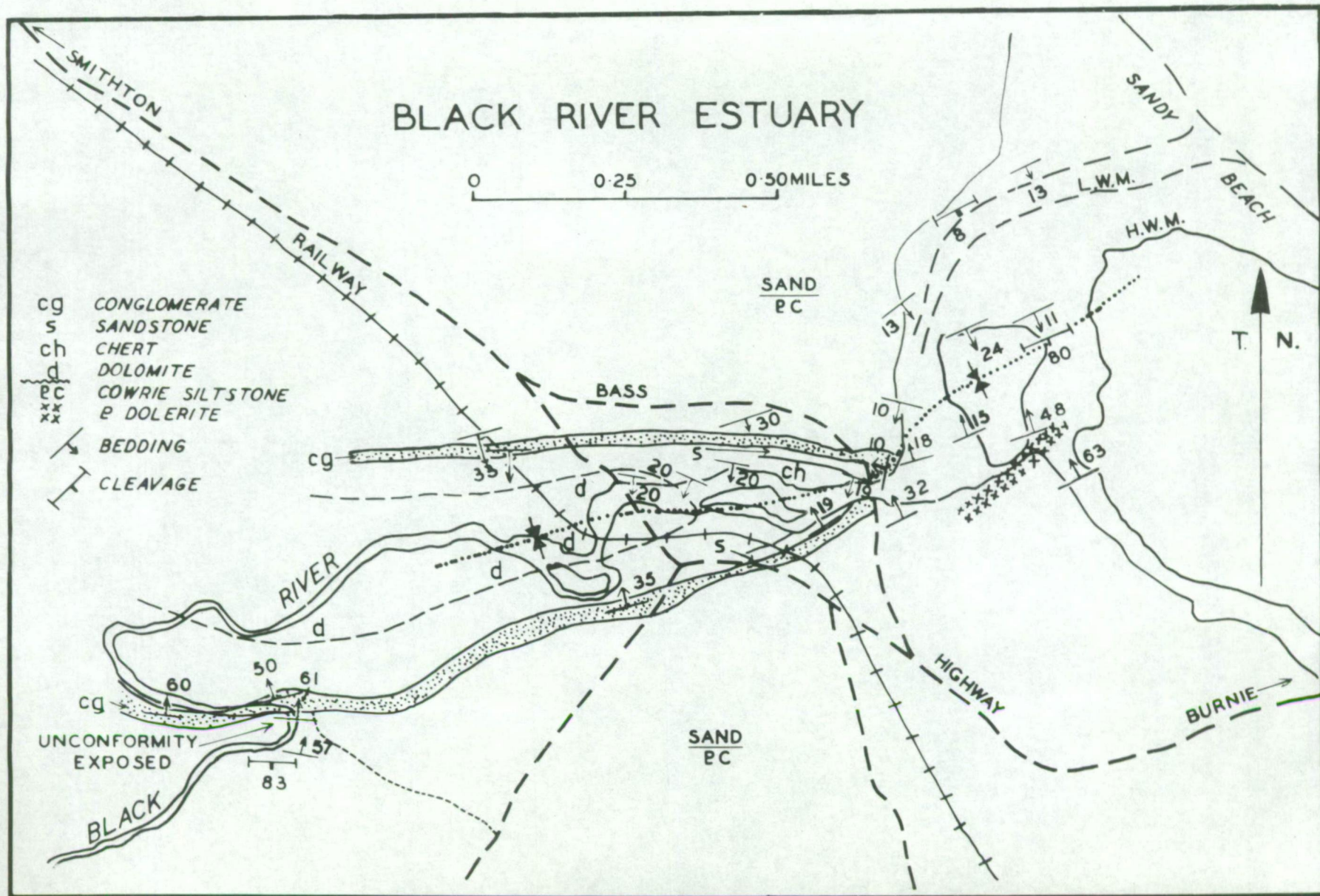
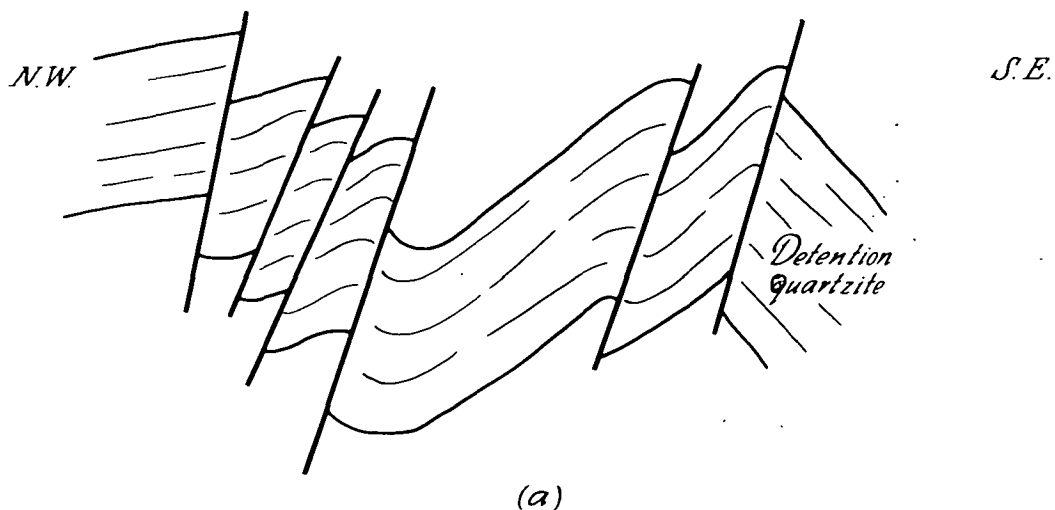


Figure 17.

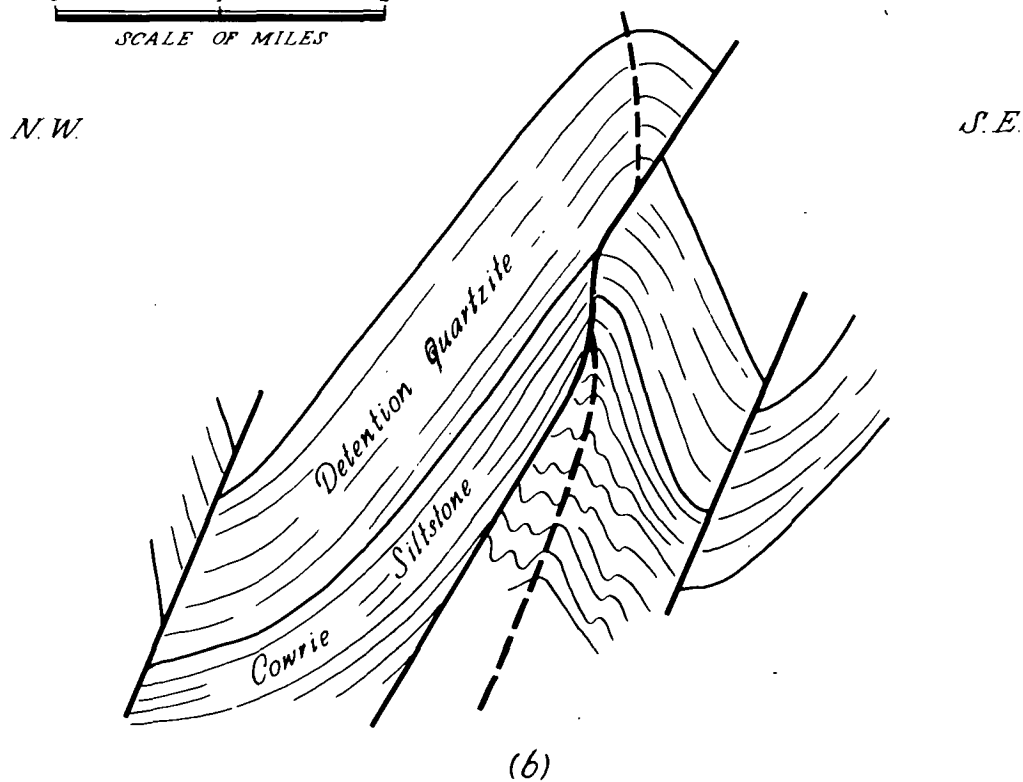
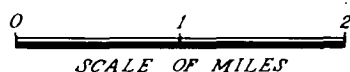
The Detention Sub-group, between Rocky Cape and Sisters Beach is broken by a series of high-angle thrusts which followed the initial folding, (Figure 18-a). The direction of movement on these faults is assumed to lie in the fault plane, normal to the axis of rotation of the associated folds. As a result, the base of the Irby Siltstone, which strikes out to sea, is progressively stepped inland and appears as remnants in fault wedges and synclinal troughs. The fold axes pitch 40° northeast in the axial planes, which generally strike northeast-southwest and dip 60° - 80° northwest.

Figure 18-b is a profile of the Sisters Hills anticline constructed from the 1 in. : 1 mile map of Table Cape (Figure 59). This anticline, and the flanking synclines are broken by high-angle, axial-plane break thrusts, all showing a west-side-up movement. The break thrust in the anticline has developed during an intermediate stage in its development. Initially a sharp break occurred in the crest in the competent Detention Sub-group and was followed by movement of about 3,500 feet.

This was taken up by bedding plane slip in the Cowrie Siltstone in the core. Further deformation proceeded by simultaneous slipping on the thrust and crumpling in the siltstone and shale. The siltstone on the northwest side of the fault is planar and without



PROFILE SHOWING REGIONAL STYLE OF
BREAK-THRUSTS, BETWEEN ROCKY CAPE &
SISTERS BEACH. (LOOKING DOWN FOLD AXES TO N.E.)



PROFILE OF SISTERS HILLS ANTICLINE
(LOOKING DOWN FOLD AXIS TO N.E.)

minor folds, whereas on the southeast side there are abundant minor folds with a conspicuous slaty cleavage.

Bedding-plane slip is essential for this style of folding in the Rocky Cape Group, yet concentric shears are not obvious. There are several lines of evidence to suggest that bedding-plane slip occurred on relatively few but discrete shear horizons, located by the interbedded lutite horizons.

The analysis of inclination of cross-bedding (Figure 6) shows that there has been no detectable simple shear within the sedimentation units. The inclinations of cross-bedding were recorded, as part of the palaeocurrent study, in order to determine the unit shear in the a - c deformational plane. Several works including Brett (1955), Pettijohn (1957) and Pelletier (1958) have detected variations from the mean due to concentric folding. Once the regional structure is known, it is possible to delineate sectors in which the inclinations of a specific azimuth should increase or decrease due to bedding-plane shear. However it has been shown that folding had no recordable effect on the inclination of cross-bedding.

The interbedded lutite beds appear to have acted as layers of intense shearing. Specimen 33327, from a thin lutite horizon nine inches thick, on the eastern headland of Sisters Beach, is a coarse siltstone with a mesoscopically conspicuous foliation parallel to the bedding. In thin section the foliation is due to alignment of clay matrix which wraps around the clastic quartz grains, some of which show fragmentation and granulation.

Generally, the lutite beds possess a well-developed lustrous slaty cleavage which is axial plane to drag folds and oblique to bedding. Plate 20-b is an example of this cleavage from a lutite bed in the Detention Sub-group from a Bass Highway road cutting on the western flank of the Sisters Hills anticline. In addition there is a later strain-slip cleavage. In the fine-grained laminae, the slaty cleavage is due to a perfect planar orientation of minute crystalline muscovite which wraps quartz micro-augen, testifying to some recrystallization. In the medium to fine-grained siltstone laminae, the cleavage is due to the orientation of minute isolated sericite flakes and the quartz grains are not recrystallized. The cleavage tends to be discrete sigmoidal planes along which there has been some displacement producing micro-boudins. (Plate 20-b).

The strain-slip cleavage in Plate 20-b is also developed in the well-cleaved slaty beds within orthoquartzite, at isolated localities in the Sisters Hills - Rocky Cape area. It is unrelated to any known folding and is later than the main period of folding.

In the thicker lutite horizons like the Irby Siltstone (Figure 19) and the Port Slate of the Detention Sub-group, the slaty cleavage is not so well developed but there are abundant oblique-slip faults parallel, or sub-parallel, to the axial planes of the drag folds.

Figure 20 is a map of a remnant fault wedge of the Irby Siltstone in the ob port on the eastern side of Rocky Cape. The folds have a constant sense of vergence indicating a dextral or easterly movement of the more superficial layers over the lower layers. The associated faults trend 093° and are parallel to the major thrust faults. The majority have a dextral component of about six inches.

The strain pattern envisaged for the deformation of the thick orthoquartzite-siltstone sequence is one in which the orthoquartzite has split into mechanically rigid slabs which have merely been bodily rotated. Bedding plane shear in the quartzite has been at a minimum, and most of the strain is accommodated by shear along the lutite horizons. Such a strain picture, resulting in

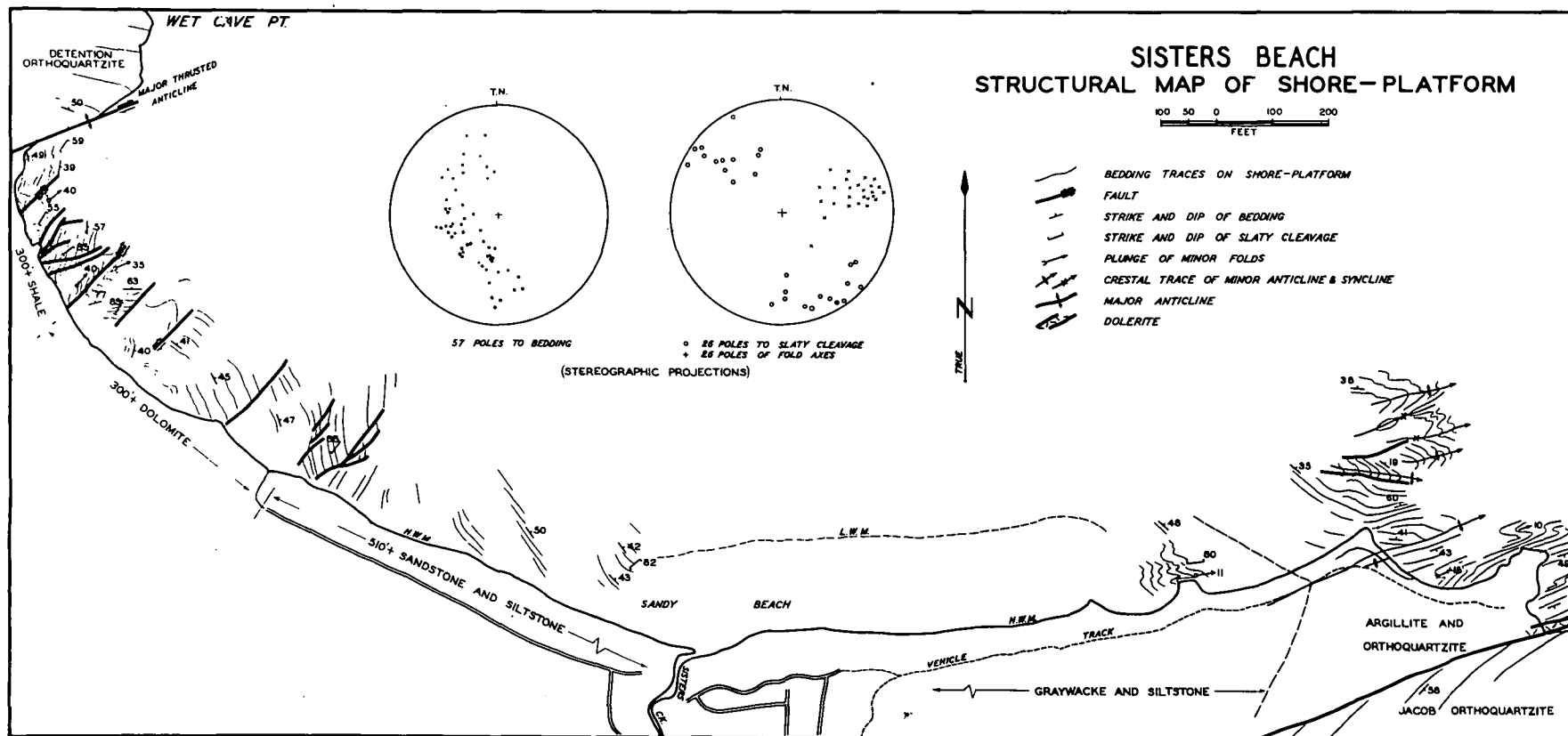


Figure 19.

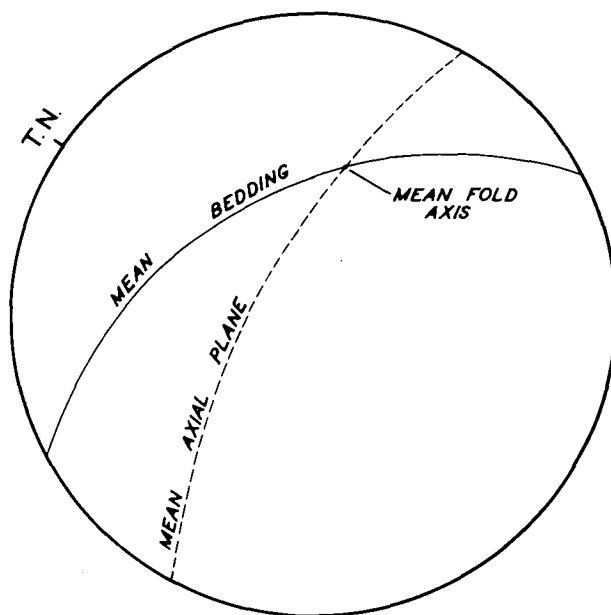
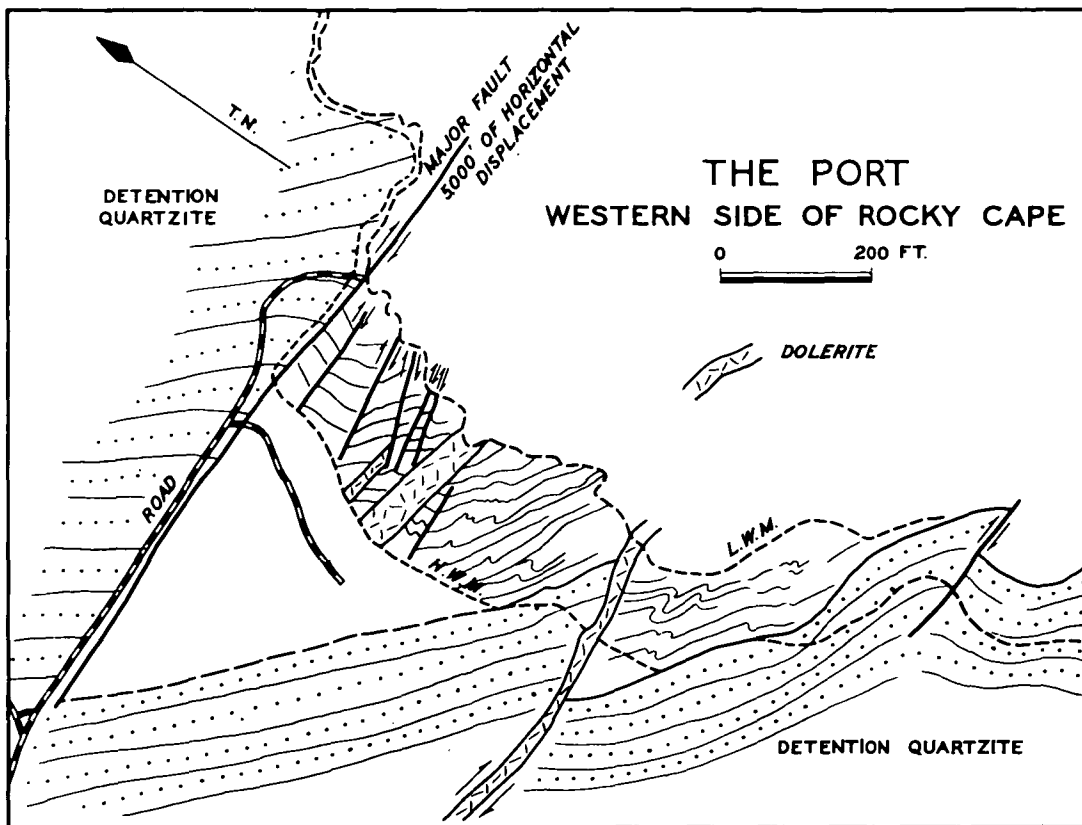
accordion-like folds allows tensional zones to develop in the axial planes of the folds, and this may have influenced the intrusion of some of the dolerite dykes.

MINOR STRUCTURES IN THE COWRIE SILTSTONE

The minor structures in the Cowrie Siltstone reveal a progressive change in style from east to west across the extent of its outcrop. This is manifest by a decrease in intensity of cleavage development accompanied by a more brittle style of deformation, and may possibly be correlated with decreasing tectonic depth at which the rocks were deformed.

Sisters Hills Area

In the core of the Sisters Hills anticline, exposed in the Sisters Hills railway cuttings, the folds are cylindroidal with a fan cleavage in the siltstone and an axial-plane slaty cleavage in the mudstone. Although the laminated rocks split more readily along bedding than cleavage, in thin section the cleavage is seen to penetrate down to the individual quartz grains. The cleavage is due to anastomosing shreds of brown-stained sericite which define lens-shaped slices. The thickness of the slices is the order of the largest quartz grain. The interstitial clay matrix has only a weak preferred orientation in the cleavage. Generally, the large detrital muscovite



STRUCTURE OF AN INFAULKED WEDGE OF IRBY SILTSTONE
IN 'THE PORT', WESTERN SIDE OF ROCKY CAPE

Figure 20.

is without preferred orientation, and the quartz grains are the original unmodified detrital grains showing no sign of fracturing or granulation.

The detrital muscovite flakes in some of the more severely deformed siltstones are, however, oriented in the cleavage. For example, Specimen 33254 (Plate 20-a) is a coarse siltstone from the shear zone in the core of the Sisters Hills anticline. At this locality there are abundant tight isoclinal folds. The muscovite (up to 2 mm in length) is large by comparison with the mean grain size (0.07mm) of the quartz, and since the bedding is perpendicular to the cleavage, the detrital muscovite has been rotated through large angles. The flakes show slight but clear evidence of bending and splintering but have not been recrystallized. The cleavage in the rock is due to closely spaced anastomosing stringy aggregates of sericite, and to a preferred orientation of the fine crystalline sericite matrix. Once again, there is no fracturing or granulation of the original clastic quartz grains. There is thus no textural evidence of recrystallization associated with cleavage development apart from the orientation of the sericite.

The purely mechanical rotation of the detrital muscovite, which is up to 50 times the size of the

quartz grains, is indicative of great mobility of the rock, in which intergranular bonding and friction was at a minimum.

These siltstones contain up to 40% matrix so it appears that the rock deformed by flowage of the matrix, thus cushioning the quartz grains which at all times behave rigidly and passively. This suggests that the rock, although buried under at least 17,000 feet of sediment, was uncompacted at the time of deformation and mobility may have been achieved by the generation of high internal pore pressures.

Cowrie Point Area

On the foreshore between Detention River and Black River, are many minor folds which are related to the main east-northeasterly trending folds. The orientation data for bedding, slaty cleavage and fold axes are given in Figure 21.

The folds are invariably coupled, (see Chapter 10 for definition of "coupled fold"), with a syncline immediately southeast of an anticline, indicating a major anticline to the southeast. A major anticline trending 080° intersects the coastline two miles west of the mouth of the Detention River. The folds are open concentric, disharmonic and non-cylindroidal, with half-wavelengths from 5 to 50 feet.

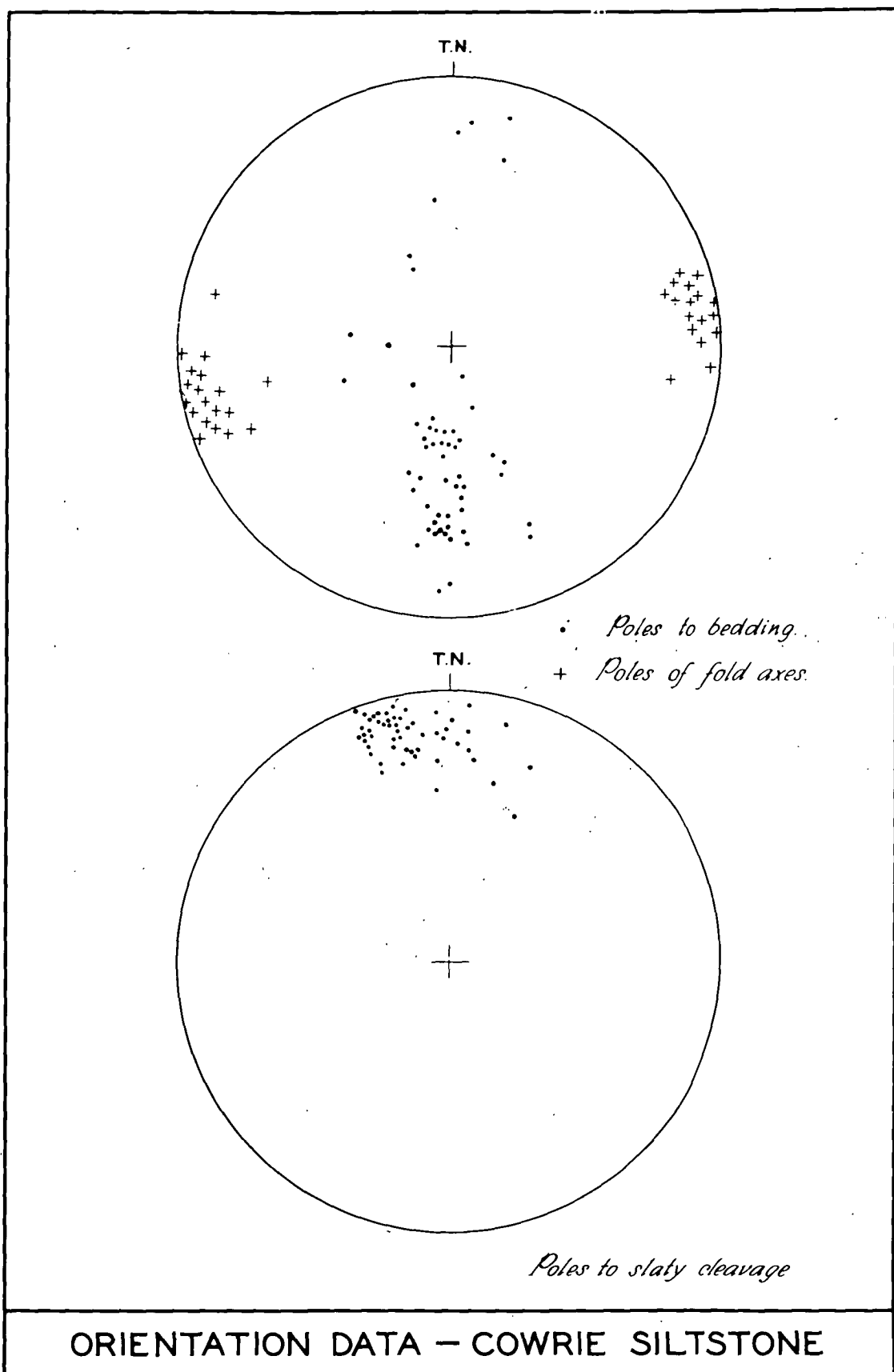


Figure 21.

The folds are disharmonic in profile and plan, and disappear along the axial-plane trace by becoming smaller and tighter, or larger and more open. Frequently they die out against faults slightly oblique to the axial plane. Folds en échelon occur, in which each fold is doubly plunging, and one fold crest has the opposite plunge to its coupled partner. A master joint is commonly present, or a longitudinal crest fault along which a differential scissors-type movement has occurred.

The cleavage also reflects the change to a more brittle style of deformation. Between Crayfish Creek and Black River, a weak slaty cleavage is found only in the mudstone and shale. Here, there is very little preferred orientation of the clay matrix, and the cleavage is defined by minute anastomosing shreds of sericite.

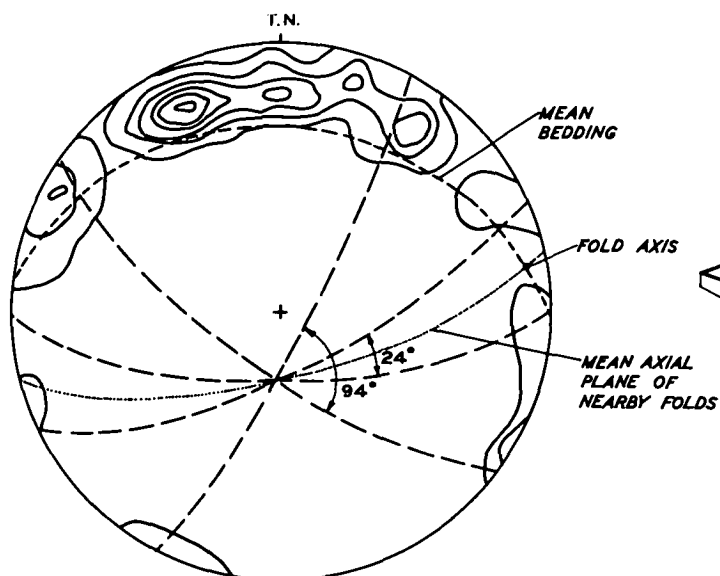
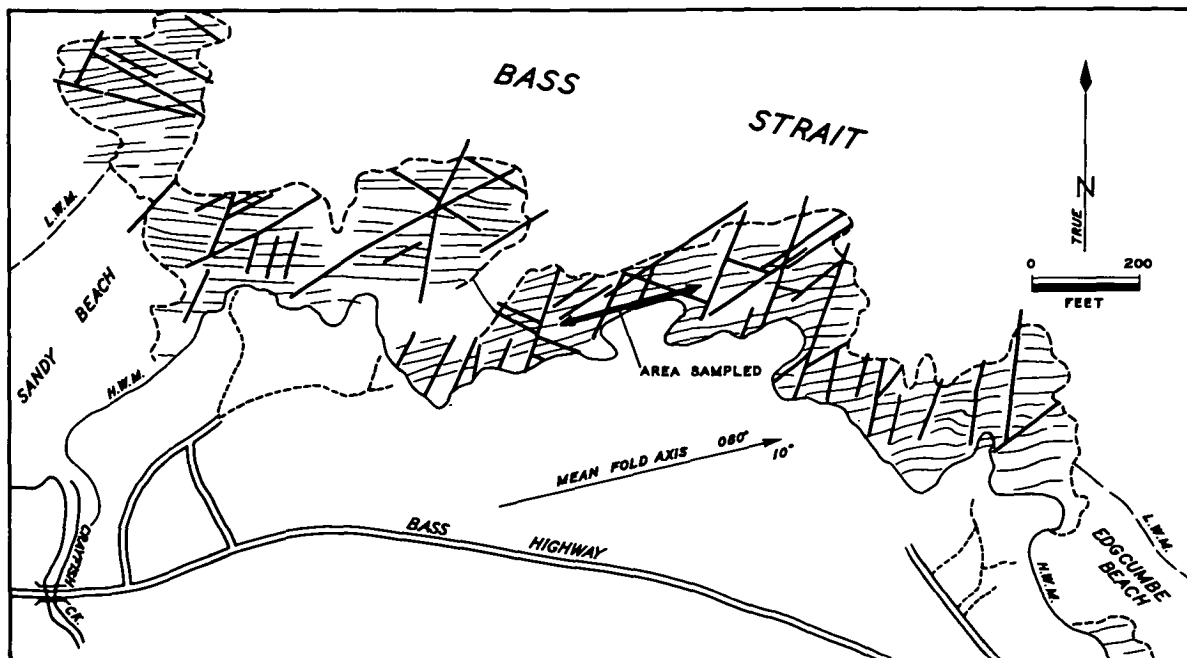
The medium- to coarse-grained siltstone beds are without cleavage, but have a set of closely spaced regular joints spread symmetrically about the fold axis. When viewed in profile it resembles fracture cleavage, but when seen on the bedding surface, they are acutely intersecting planar joints. The siltstone thus splits into long slender prisms of triangular or rhombic cross section. The joints are spaced from $\frac{1}{4}$ to 2 inches apart. The joint surfaces are smooth and perfectly planar, without plumose structures, slickensides or displacements

at intersections, and on form alone there is no firm indication whether they are shear or tension joints.

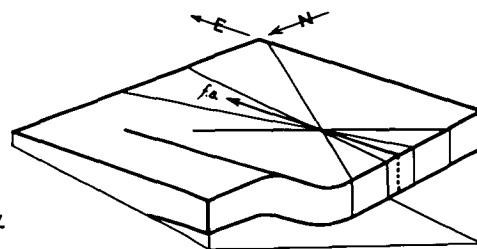
There is a specific relationship between the orientation of joints, faults, master joints and folds, suggesting synchronous development under low confining pressure. Figure 22 is a map of the shore platform between Crayfish Creek and Edgcumbe Beach showing the outcrop traces of the bedding surfaces and the master joints. The master joints form sets trending 028° , 125° , and a third which is more variable between 058° - 076° .

A total of 100 poles to joints taken from the area indicated, are plotted in Figure 22. These have been sampled from five stations regularly spaced along 100 feet of the one siltstone bed. Also plotted on Figure 22 is the mean attitude of bedding and the weak slaty cleavage in the mudstone, the intersection of which gives the statistical fold axis. All the joints lie in a great-circle girdle whose axis lies in the fold axial plane and exactly 90° from the fold axis. This is one of the symmetry axes of the fold system, being the intersection of the axial plane, and the a - c fabric plane. The mean axial plane lies at about 70° to the bedding, so that the girdle of joint poles does not lie exactly in the bedding.

MASTER JOINT PATTERN AT CRAYFISH CREEK



100 POLES TO JOINTS
CONTOURS 1-5-10-15-20-25%
EQUAL AREA



RELATIONS BETWEEN
FOLDS AND JOINTS

Figure 22.

There are four maxima which define four dominant joint directions (i.e. $015^{\circ}/82^{\circ}\text{SE}$, $052^{\circ}/72^{\circ}\text{SE}$, $076^{\circ}/70^{\circ}\text{SE}$ and $112^{\circ}/75^{\circ}\text{S.}$), all of which correspond with the master joint sets. These maxima form two sets of paired joints, both symmetrical about the fold axis. One set has a small dihedral angle of 24° so that each plane is inclined at 12° to the axial plane. The other set has a dihedral angle of 94° with the plane inclined at 47° to the axial plane.

The jointing is symmetrically related to the folding, and must be related to the stress field producing the folding. However it is unrelated to any locally induced folding strain because the jointing is equally well-developed in mesoscopically folded domains and regions of planar bedding.

Conjugate symmetrical joint sets are usually interpreted according to the Mohr-Coulomb theory of shear failure, in which the planes of failure are planes of high shear stress inclined at angles of less than 45° (usually about 30° , Hubbert, 1951) to the direction of maximum principal stress. The maximum principal stress thus bisects the acute dihedral angle. This theory is inadequate for the joints in the Cowrie Siltstone because the dihedral angle is, in one set too large, and in the other set too small.

In order to explain joint sets of small dihedral angle, Muehlberger (1961) proposed a modification of the Coulomb theory of failure, in which the coefficient of internal friction increases at low confining pressure. The Mohr circles will meet the Mohr envelope at smaller angles for lower stress values. The principal maximum stress then bisects conjugate joints of decreasing dihedral angle, until an extension fracture forms parallel to the principal maximum stress. According to this theory, the least principal stress must pass into the tensional field, and so the jointing records the very first (or the very last) stress in a single act of deformation. Muehlberger cites examples of a single set of fractures having the features of extension fractures, which pass into acutely intersecting conjugate sets of small dihedral angle.

As stated, Muehlberger's proposal is not readily applicable to the jointing in the Cowrie Siltstone. It does not readily explain the other conjugate set with a larger dihedral angle, and also does not explain why the acute bisectrix, which should be taken as the maximum principal stress, is directed along the fold axis. It is clear from the near-by folds that the maximum principal stress operating

during folding must have been oriented at right angles to the fold axis, and probably specified more precisely as being normal to the mean axial plane.

The conjugate set with the larger dihedral angle (of 84°) has its acute bisectrix in this direction and is therefore directly related to the folding. Under this stress field, the intermediate principal stress, given by the axis of the joint girdle, would be vertical and equal to the overburden load. The joint set of small dihedral angle suggests a decline of the principal maximum stress to a point where it becomes the minimum stress, and actually becomes tensile. This may be achieved by a relaxation of the stress at the end of the act of folding, or by local tensile zones related to local fold perturbations in the competent siltstones during folding. There is no clearly consistent order of development in the form of terminations of one set or another to choose between these two cases.

MAJOR EAST-WEST FAULT

A major fault, expressed as a line of structural mis-match, extends from Boat Harbour westward to the Black River, (Figure 16). Matching of traces and oppositely dipping limbs of the major folds, indicates a transcurrent movement with a north-side-east movement of 5 miles. Thus, the

orthoquartzite ridge on the western limb of the Newhaven syncline joins the orthoquartzite (Detention Sub-group) forming the line of hills between Rocky Cape and Sisters Beach. The varvoid siltstone in the vicinity of Mawbanna beneath the orthoquartzite becomes the Cowrie Siltstone which is beneath the orthoquartzite at Rocky Cape. The Irby Siltstone at Sisters Beach is displaced to the west to become the siltstone in the core of the Newhaven syncline at Montumana.

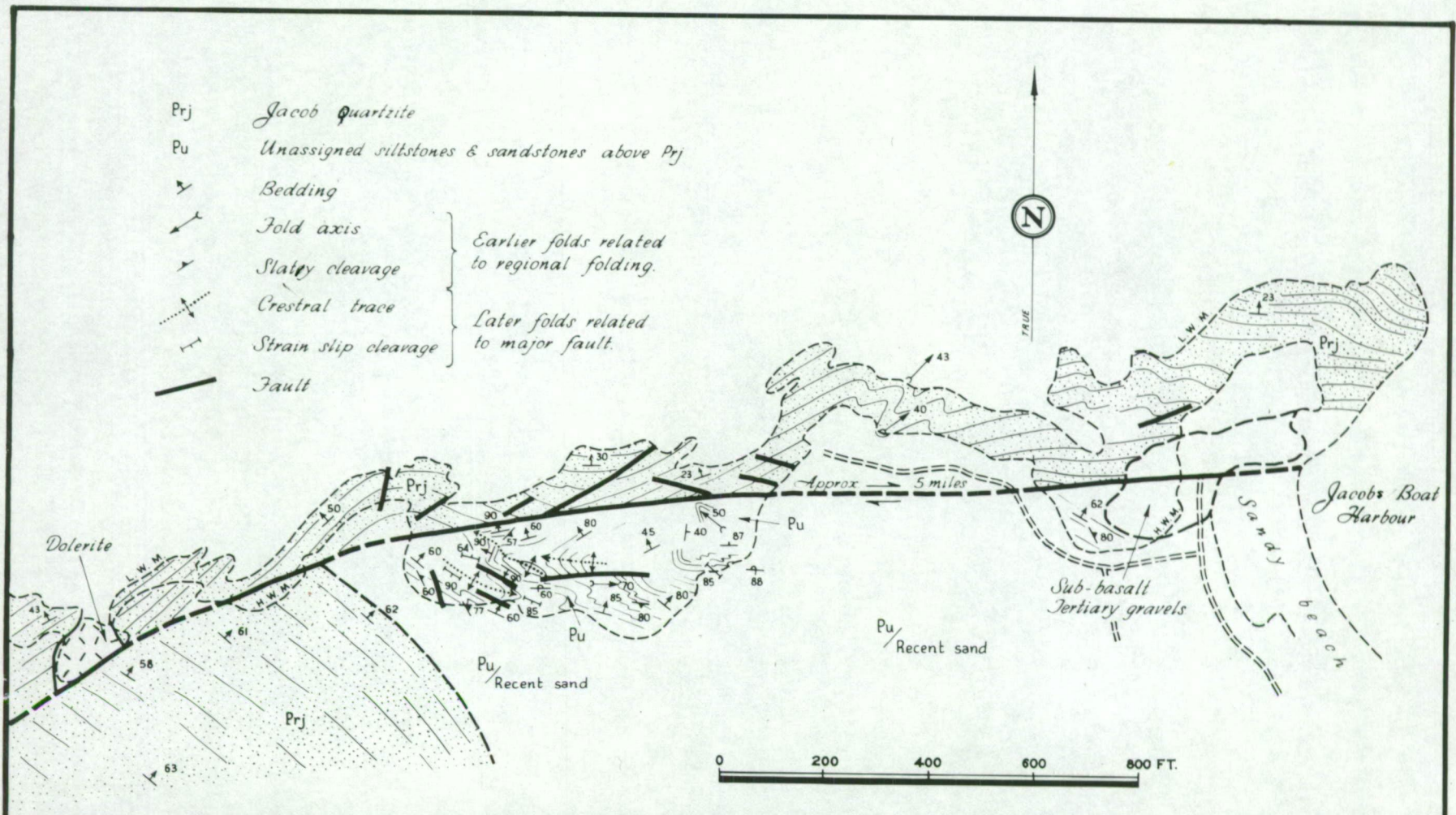
The syncline in the Jacob Quartzite west of Jacobs Boat Harbour then becomes the correlate of the Newhaven syncline. This fold is cut by the fault, but the Jacob Orthoquartzite does not appear in the core of the Newhaven syncline near Montumana as would be expected. This suggests that there has been a vertical component of south-side-up, to allow the remnant of orthoquartzite to be eroded away. Because the syncline plunges 30° - 40° northeast, the amount of vertical movement need only be small (about 500 feet) by comparison with the transcurrent movement. The Sisters Hills anticline is cut and displaced out to sea to the east of Jacobs Boat Harbour. The broad curvature of the bedding on the headland may indicate the proximity of this anticlinal trace.

Complex re-folding occurs where siltstone is involved in the fault zone. This is observed in

the Sisters Beach road-cuttings, $1\frac{1}{2}$ miles from the beach, and on the shore platform immediately west of Jacobs Boat Harbour. Here, the line of the fault can be traced for 2,000 feet along the shore platform, and the actual fault plane is observed in several places. No slickensides have been observed.

Figure 23 is a map of the fault zone at Jacobs Boat Harbour. The crumpling is restricted to the siltstone (un-named, above the Jacob Quartzite) and takes the form of en échelon, doubly-plunging anticlinal domes with a strain-slip axial-plane cleavage. These structures refold the pre-existing minor folds and cleavage which are related to the regional folding. In addition, there is a set of small-scale vertical shear planes trending west-northwest which are expressed as axial planes of angular chevron folds in the siltstone, and small transcurrent faults in the quartzite. These have a sinistral movement and may therefore be conjugate faults related to the dextral movement of the major fault.

The age of this fault is uncertain. It post-dates the regional folding in the Rocky Cape Group and is earlier than the Tertiary basalt. It does not fit in with the general structural plan of anything that is known about the tectonics of the



MAJOR EAST-WEST FAULT AT JACOB BOAT HARBOUR

Figure 23.

northwest coast of Tasmania. It is not a geomorphological feature and the associated crumpling distinguishes it from the Mesozoic or Cainozoic epj@rogenic faults found elsewhere in Tasmania. Extrapolating further to the west into the Smithton area, it does not seem to affect the large north-south Cambrian fault mapped by Gulline (1959) and Carey and Scott (1952). It is sugges_ted that this east-west transcurrent fault is a Precambrian structure. A smaller, east-west fault, which appears to have a transcurrent component, off-sets the major structure in a dextral sense in the vicinity of Milabena (Figure 16).

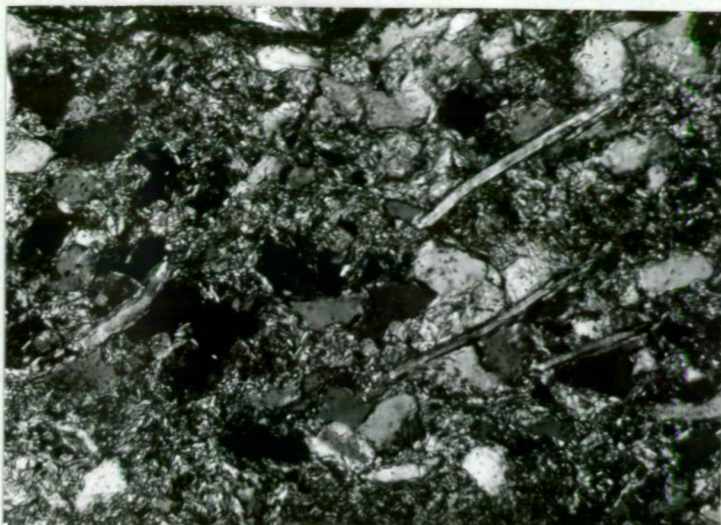


Plate 20a Photomicrograph of cleavage in coarser variant of Cowrie Siltstone, core of Sisters Hills anticline, Sisters Hills. Cleavage defined by alignment of mica flakes and is at right-angles to bedding. Sp. 33254, x 75.

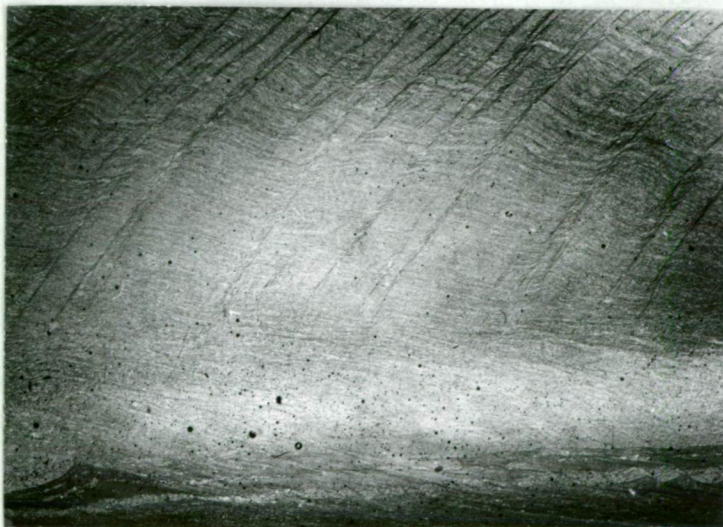


Plate 20b Photomicrograph of slaty horizon in silty layer, Detention Sub-group, Sisters Hills. Slaty cleavage is acute to bedding and is crenulated by later strain-slip cleavage. Sp. 33267, x 2.

C H A P T E R 10STRUCTURAL ANALYSIS OF THE BURNIE FORMATIONINTRODUCTION

Despite the excellence of outcrop, unravelling of the major structure of the Burnie Formation is confronted with several difficulties. The formation is a well-bedded monotonous alternation of arenite and mudstone, a sequence favourable for the development of abundant mesoscopic folds. However, by the very nature of the rock type, there are no marker horizons which can be used to determine the major structure. Normal procedures whereby the major structure is deduced from the stratigraphy cannot be followed. In fact, the reverse is true and any attempt at stratigraphic subdivision must follow after the major structure is known.

The structure is complex on all scales due to repeated deformation involving five phases of deformation. Structural analysis is made more difficult because of some peculiarities of superposition. The individual phases are nearly all coaxial, and the mesoscopic folds of one generation occur in zones which are spatially segregated from zones of other generations. Problems in differentiating between fold phases arise because

the areas of overlap of these zones are small. Thus, examples of refolded structures are uncommon and a false impression is given of rapid changes in style of the one fold generation. A purely statistical approach to the macroscopic analysis of mesoscopic structures, as envisaged by Dahlstrom (1952) and Turner & Weiss (1963, p.146-163) is not fruitful and may be misleading. As well as orientation data of bedding, cleavages, axial planes and fold axes, it is essential to record other attributes such as style, symmetry, vergence, facing and the orientation of the enveloping surface of the minor structures, with the object of delineating zones of like attributes.

METHOD

The major structure has been deduced by continuous detailed mapping on the mesoscopic scale across the outcrop belt of the Burnie Formation, and piecing together the minor structures. The detailed study has been confined to the shore platform between Penguin and Doctors Rocks (4 miles east of Wynyard). At low tide, up to 200 yards of shore platform is exposed, much of which is continuous exposure. However there are some long breaks in the exposure due to coverings of Tertiary basalt and Recent beach sands, for example, between

Penguin and Sulphur Creek, and also between Wivenhoe and Burnie.

Mapping was done directly onto large-scale aerial photographs on scales ranging from 100 ft : 1 in to 250 ft : 1 in. All structures that can be represented on this scale were mapped, and this includes almost every minor fold visible on the shore platform, as well as the faults. The folds have been mapped as the actual bedding-plane trace on the shore platform. Representative attitudes of all structural elements related to all the fold phases were measured.

The geology was transferred from the aerial photographs onto a base map of scale 200 ft. : 1 in., by means of an epidiascope. These base maps were originally prepared for the Bass Highway survey by the Public Works Department, and are the result of accurate surveying. Some slight distortion of the structure results in transferring the data from photograph to map, but there is no cumulative error.

These maps have been reduced to a scale of 400 ft. : 1 in. for presentation in this thesis (Figures 54, 55, 57 and 58). This chapter should be read in conjunction with these maps. The lines that represent folds are actual bedding traces that have been followed out in the field, and only in

structurally complex zones are they diagrammatic. The spacing of these lines is arbitrary, depending on the complexity of the structure and the extent of bedrock exposure, and is not intended to represent particular features of the stratification.

TERMINOLOGY

Phase of Deformation

A phase of folding is a time division of an orogeny to which can be related a distinct act of folding or other penetrative deformation. At any one point within the deformed belt there will be a succession of movements which are termed the phases of that orogeny. The succession may be established by dating of regionally significant unconformities by palaeontological methods, or by demonstration of superposition by structural analysis. No specific interval of time between the events is implied. It is important to note that the phase is defined with respect to the recorded strains in the rock, rather than the presumed tectonic stresses which would be expected to vary continuously with time.

The various phases are designated in chronological order, P1, P2, P3 and so on. This terminology refers to the area as a whole and not to the sequence at any one point. Thus, because

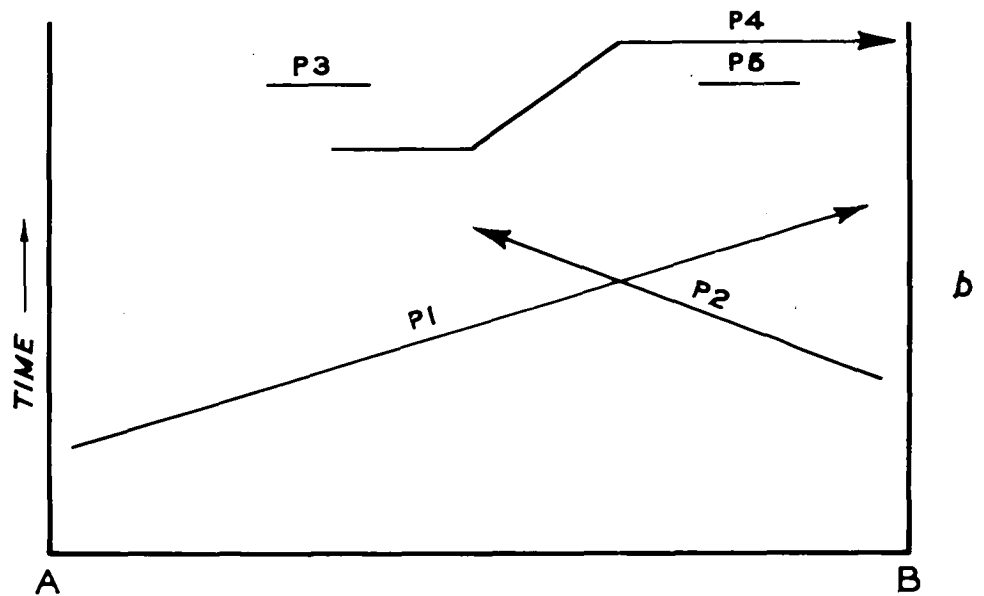
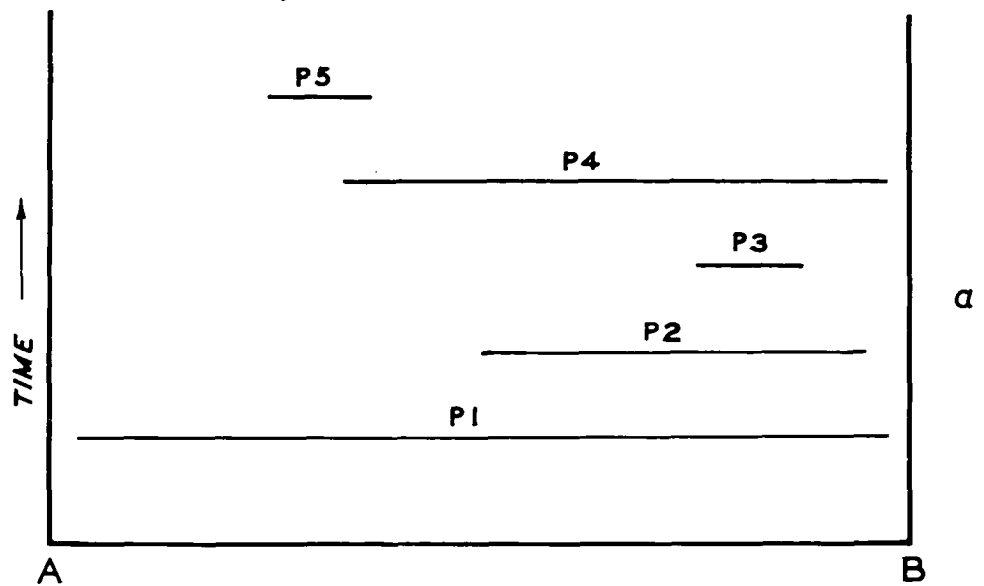
of the segregated nature of the phases, the situation can arise where the P2 phase may be the first deformation at one locality, but at another locality that same phase is later than the P1 phase. The usage of some writers (Wood 1963) of classifying the folds as P1, P2, P3 etc. according to style alone, without any implication of a sequence of deformation is not followed.

The methods and problems of correlation of phases are somewhat analogous to stratigraphic correlation. The most reliable method is tracing the lateral continuity of a group of folds from one area to the other, a procedure that is analogous to tracing the lateral continuity of a stratigraphic horizon. Correlation can further be made on the grounds of identity of character of a group of structures. The character of a group of related structures is expressed in terms of its style, and orientation with respect to external geographic co-ordinates. This is analogous to correlation by lithologic identity, and the problems analogous to those arising from facies changes are also present. Correlation can also be made by the position in the structural sequence which is analogous to position in the stratigraphic sequence. The obvious problem here is strong spatial segregation of the individual fold phases,

so that various events in the sequence are missing.

These three categories of correlation have all been used in the structural analysis of the Burnie Formation, although the method of tracing lateral continuity has been used in preference. It is therefore felt that the correlations of structures from one area to another are valid.

This method of designating and correlating deformation phases assumes a simple superposition of individual phases which were formed more or less at the same time. A hypothetical sequence is shown diagrammatically in Figure 24-a. If, however, the individual phases are markedly diachronous, correlation becomes problematical. It is possible that deformation could commence at one point and slowly propagate its way across a deforming basin. At the same time a phase of deformation having a different style and orientation may form on the other side of the basin and propagate in the other direction. It is then possible that the two paths could intersect. Even more complex structural sequences can result from the intersection of phases with different propagation velocities. The same hypothetical sequence in Figure 24-a could then be represented as in Figure 24-b.



DISTANCE ACROSS DEFORMED BASIN

SEQUENCES IN DIACHRONOUS (b)
AND NON-DIACHRONOUS (a)
FOLD PHASES

Figure 24.

Such diachronous phases may be revealed by anomalous sequences after intensive structural analysis. One other line of approach is that outlined by Rickard (1965) whereby muscovite flakes defining cleavages associated with individual phases are dated radiometrically. No anomalous sequences are known from the Burnie Formation.

Coupled Fold

This is a fold consisting of an anticlinal hinge joined by a common limb to an oppositely closing synclinal hinge. The coupled folds are usually separated from each other by long straight planar outer limbs. In this respect they are similar to "drag folds", but this term is avoided because of its dynamic implications. Coupled folds may occur singly or in groups.

Facing

Facing can refer both to attitude of bedding and folds. The facing of the bedding after tilting (Shrock 1936, p.18) is the direction in which a pole to the younger surface is now pointing. It is determined most easily in these rocks by small-scale cross-bedding, and also less readily by load cast structures, graded bedding, and sole marks on the arenite beds.

Shackleton (1957, p.363) extended the term to cover folds. A fold is said to face in the

direction normal to its axis, along the axial plane and toward the younger beds, thus coinciding with the facing of the beds at the hinge. This is a particularly useful term in areas of regional overturning to describe simply the fold orientation, and it also has some kinematic implication. The direction of facing of a large recumbent fold is also the direction of tectonic transport, or the direction of movement of the more superficial layers of the earth's crust.

Enveloping Surface

For deducing the macroscopic structure from a set of mesoscopic structures, the enveloping surface is a useful concept. The enveloping surface is the imaginary surface that can be drawn tangentially to the crests of successive antiforms (or synforms) developed on a single folded surface. Its orientation can be deduced from the mean B-axis and the trace of successive folds on the shore platform, and will give the gross orientation of the bedding.

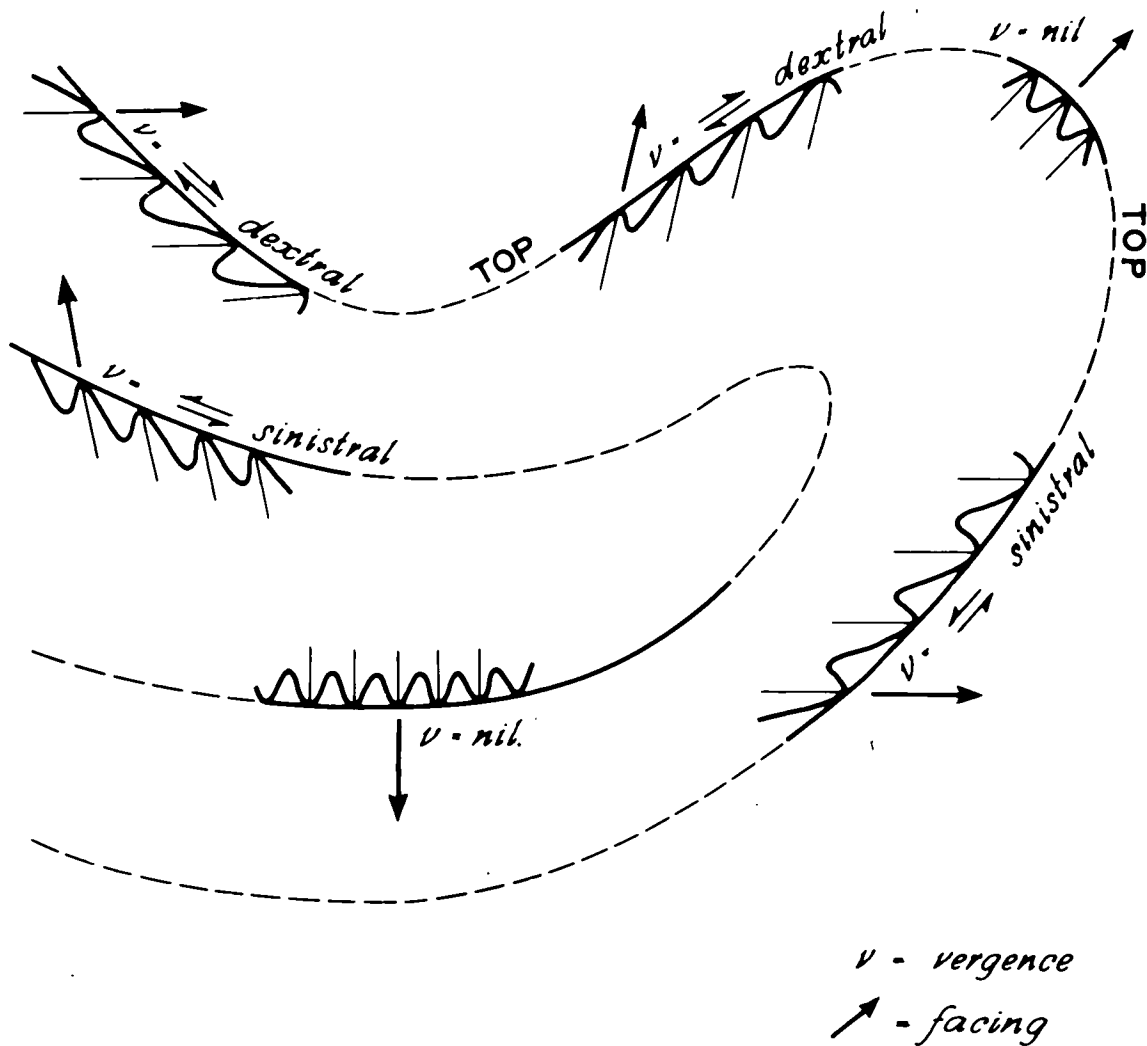
Vergence

The relationship between the enveloping surface and the axial planes of a group of folds determines its symmetry. When the axial planes are normal to the enveloping surface the folds are bilaterally symmetrical and have orthorhombic symmetry.

When inclined, the folds become bilaterally asymmetrical with monoclinic symmetry. The vergence refers to the sense of rotation of the axial planes away from the perpendicular intersection with the enveloping surface as the folds become progressively more asymmetrical.

H. Cloos (1936) appears to have been the first to use the term vergence to indicate the direction of overturning of minor folds. Ramsay (1958, p.286), Sturt (1961, P.142), Harris and Rast (1961, p.51) and Wood (1963) have also used the term in the same context, and applied the words sinistral and dextral.

It is difficult to express vergence in words. It can be designated sinistral (anticlockwise) or dextral (clockwise) as implied by the sense of rotation of the asymmetry of the folded surface. These terms can be useful for plunging folds, but it should be noted that a dextral vergence looking down the axis becomes a sinistral vergence when looking up the axis. In the areas of low plunge, the vergence can be specified by a direction, i.e. westward or eastward, according to the apparent direction of overturning implied by the asymmetry. The easiest way of specifying vergence is diagrammatically, and this is the procedure adopted in this chapter.



RELATIONS BETWEEN FACING, VERGENCE AND ENVELOPING SURFACE

Figure 25.

Vergence and facing are different in concept and are independent of each other. Facing helps specify the orientation of a group of folds, whereas vergence is a symmetry concept. A group of mesoscopic folds can face in the opposite direction to the vergence. Likewise, a group of mesoscopic folds can have any inclined facing yet have no vergence, as in the core of a recumbent fold. These relationships are shown diagrammatically in Figure 25.

The use of the vergence concept involves the mapping of belts of like vergence and vergence boundaries. It can be applied irrespective of whether the mesoscopic and the macroscopic folds are of the same generation. Even in areas of poor outcrop the vergence can be determined by the relations of bedding to cleavage.

STRUCTURAL BELTS

From the detailed maps it is possible to delineate regional zones that are broadly homogeneous with respect to some fundamental property, such as the orientation and facing of the enveloping surface, and the vergence of the first generation (P1) mesoscopic folds. Five of these structural belts can be recognised, and for ease in description they are given formal names. From west to east they are as follows:-

Table 6

SOMERSET OVERTURNED BELT	COOEE HINGE	PARKLANDS FLAT BELT	ROUND HILL HINGE	BLYTHE OVERTURNED BELT	PENGUIN REVERSAL	DEFORMATION
RECUMBENT P ₁ CORES ARE ABUNDANT	ABUNDANT P ₁ FOLD CORES	SCATTERED MINOR P ₁ FOLDS	OCCASIONAL P ₁ CORES	ABUNDANT P ₁ FOLDS	P ₁ FOLDS	P ₁
			PLANAR STRAIN- SLIP CLEAVAGE CUTTING ACROSS P ₁ CORES			P ₂
	SOME P ₃ FOLDS	ABUNDANT P ₃ FOLDS. EXAMPLES OF REFOLDINGS OF P ₁ BY P ₃ AT COOEE	ABUNDANT P ₃ FOLDS. EXCELLENT EXAMPLES OF REFOLDING OF P ₂ BY P ₃			P ₃
ABUNDANT P ₄ FOLDS WITH STRAIN-SLIP CLEAVAGE. CLEAR EVIDENCE OF SUPER POSITION	WEAK P ₄ STRAIN- SLIP CLEAVAGE					P ₄
				LOCAL P ₅ FOLDS CLEARLY REFOLDING P ₁ STRUCTURES. NO OTHER EVIDENCE OF AGE		P ₅
				ABUNDANT KINKS	ISOCLINAL FOLDS REFOLDING P ₁ . REGIONAL TABBER- ABBERAN FOLDS. ABUNDANT KINKS	LATER DEFORMATIONS
SUMMARY OF STRUCTURAL HISTORY OF BURNIE FORMATION						

PRECAMBRIAN DEFORMATIONS
(PENGUIN OROGENY)
MAIN

- (1) Somerset Overturned Belt (adjacent to Keith Metamorphics)
- (2) Cooe Hinge Zone
- (3) Parklands Flat Belt
- (4) Round Hill Hinge
- (5) Blythe Overturned Belt
- (6) Penguin Reversal, (adjacent to Cambrian trough).

A summary of the structures present in these belts is given in Table 6. The five phases of deformation are considered all to be Precambrian in age and due to the Penguin Orogeny. Their age is discussed in Chapter 12. At Sulphur Creek and Blythe Heads, Burns (1964) recognised two phases of folding which he designated P1 and P2. The P1 of Burns corresponds to P1 of this thesis and P2 of Burns corresponds to P5 of this thesis.

The remainder of this chapter is devoted to a description of the structural geometry of these zones, and to the description of the mesoscopic structures later than the P1 phase. The orientation data and diagrammatic summaries are presented in Figures 54, 55, 56, 57 and 58.

Somerset Overturned Belt (Figure 26)

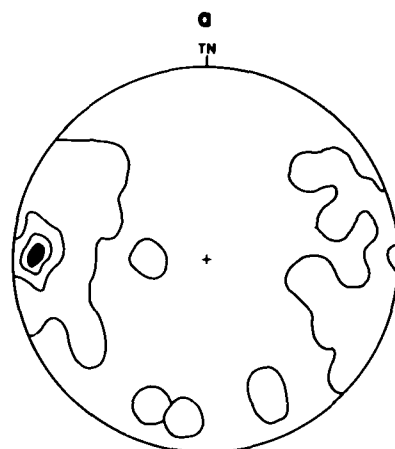
From the unconformity with the (unfolded) Permian rocks near Doctors Rocks, along to the mouth of the Cam River, the bedding is generally completely overturned. There are many P1 folds with a mean axis plunging gently toward 290°. The folds are recumbent.

with mean axial planes dipping 20° to the south (Figure 26-b). The long outer limbs of the coupled folds are overturned and the short common limbs are steep, with the bedding often re-inverted. The P1 folds face down to the south. The facings and vergence are shown diagrammatically in Figure 26-d. The enveloping surfaces of the P1 folds are approximately sub-horizontal with a dip to the south. It is difficult to specify it any more accurately on account of the P3 folding.

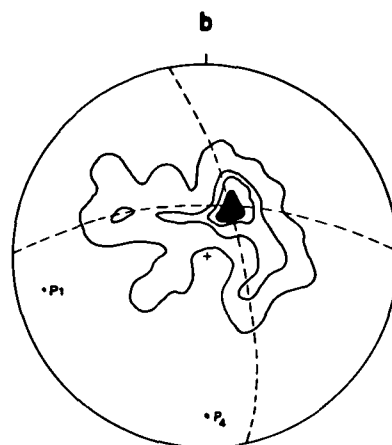
The P1 axial-plane cleavage is very strongly developed, especially west of Doctors Rocks where it is difficult to recognise the bedding. The slate is glossy, and a rudimentary microscopic lamination is developed. The arenite is also strongly cleaved and splits preferentially along cleavage rather than bedding.

This belt also contains P4 folds and an associated strain-slip cleavage. The P4 structures are most strongly developed west of Doctors Rocks, and decrease in intensity to the east. At Cam River, no P4 folds are visible and the cleavage is only sporadically developed. The orientation of the P4 folds is shown in Figure 26-c. The fold axes plunge gently to the north-northeast or south-southwest and the axial-plane cleavage is vertical.

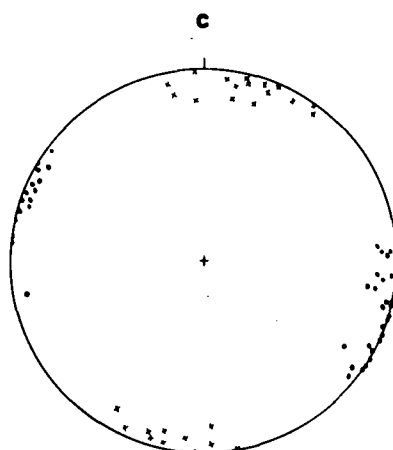
The P4 structures are characterized by concentric fold styles, axial-plane strain-slip



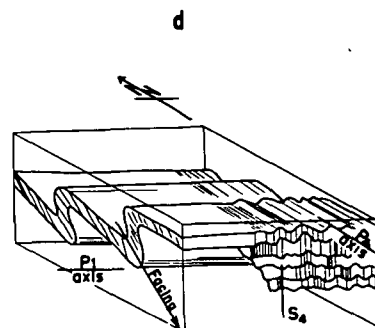
75 poles of P_1 fold axes
contours 1-3-6-9%
mean axis $258/14^\circ$



75 poles to P_1 axial surfaces
Showing position of P_1 and P_4
axes.
contours 1-3-6-9%



• 44 poles to P_4 strain-slip cleavage
• 27 poles of P_4 fold axes



Diagrammatic representation of
vergence and facing of P_1 folds.

STRUCTURAL DATA SOMERSET OVERTURNED BELT

Figure 26.

cleavage, and an axial orientation that is distinct from the other generations. In the massive or laminated slate, the folds are rounded and concentric in style with wavelengths up to 10 feet. A closely spaced strain-slip cleavage is parallel to the axial plane. Plate 21-a is a photomicrograph of this cleavage. As the slate becomes more phyllitic, the folds become sharp-crested, accordeon-type crumplings of the foliation. The folds in the finer grained beds occur as bundles of en échelon folds with discontinuous crests. These die out at the contacts with the arenite beds, giving way to broad undulating zones of concentric folds with wavelengths up to 50 feet. As a rule, bedding is passive in the arenite and folding occurs by slip on the kinematically active P₁ cleavage.

Cross-folding seems to be associated with the P₄ phase where thickly bedded and weakly cleaved arenite is involved. Some near-perfect examples of circular domes and basins are found one mile west of Somerset (Plate 21-b). Generally, however, two trends intersecting or bifurcating at angles of 45° to 90° are present. These produce triangular, heart-, or amoeboid-shaped bedding traces on the shore platform. No cleavage is associated with this cross-folding.

The cross folding is clearly not due to the interaction of P₁ and P₄ structures. The

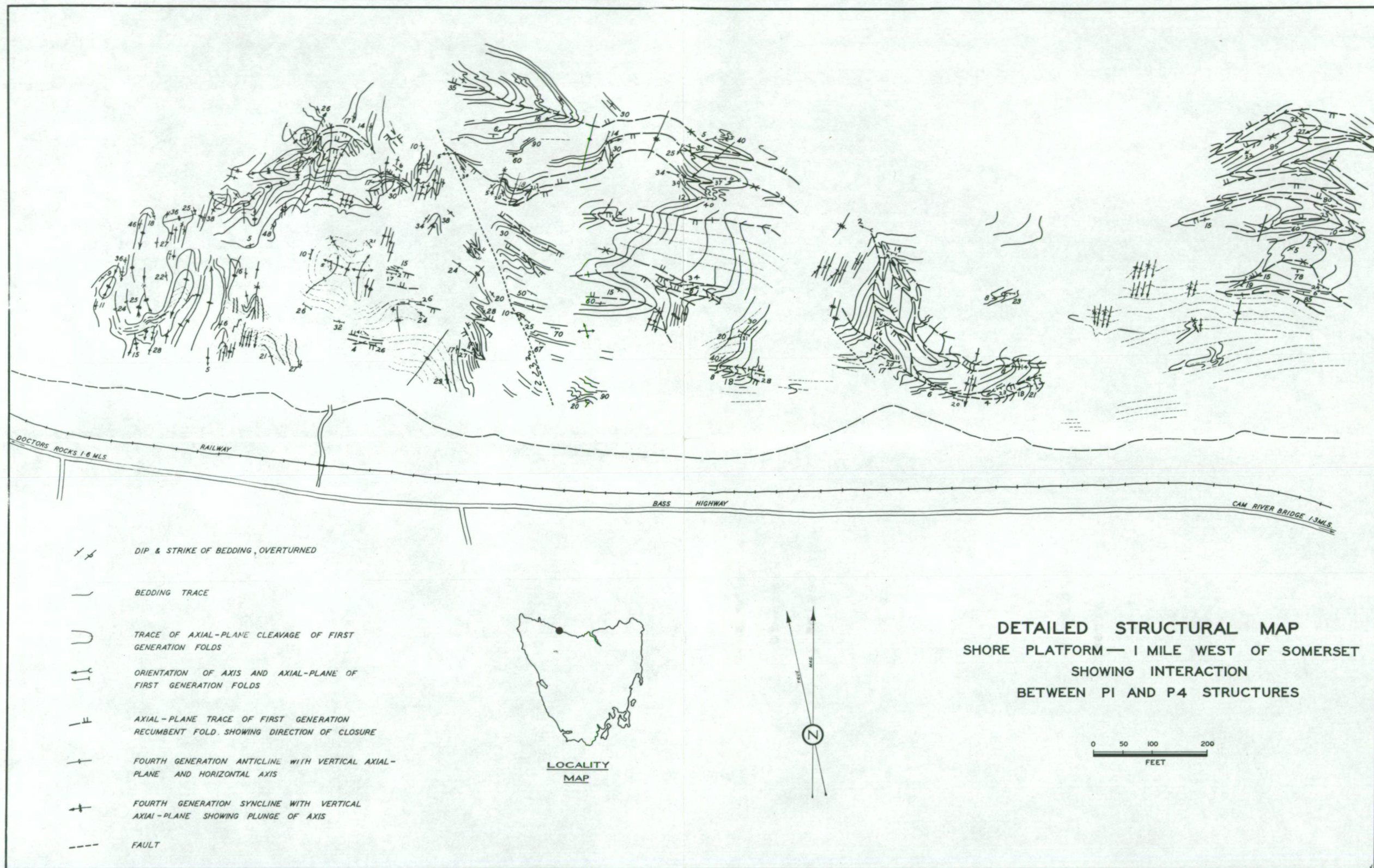


Figure 27.

fact that doming occurs only in the arenite and not in the slate, and that there are all gradations between P₄ cylindrical folds and domes, indicates that the domes and basins were formed by later synchronous cross folding during the P₄ phase rather than by the superposition of two distinct phases. Cross-folding of this type, where the domes and basins are equidimensional is a rarity in the geological literature. Hypothetical examples are described by O'Driscoll (1962) where the fold profiles are produced by vertical movements on two mutually inclined shear surfaces. However the suggested origin of O'Driscoll is not supported by the lack of cleavage and the concentric profiles. This appears to be purely a buckling phenomenon, requiring a stress field of axial or tetragonal symmetry with all compressive stresses acting in the plane of the layering being equal, and the least principal stress directed vertically.

Excellent examples of refolding of P₁ structures by the P₄ folds are found on the foreshore between Somerset and Doctors Rocks. Figure 27 is a map of minor structures of a small section of the foreshore, and shows in many places the axial traces of P₁ folds intersected and warped by the P₄ traces. In Figure 26-a is given the orientation of P₁ fold axes which have been dispersed by P₄ folds, however

since P₄ folds are only broad open undulations, the degree of dispersion is not great.

There is a general tendency in Figure 27 for the P₄ folds to be more strongly developed in zones of planar flat-lying bedding, free from earlier mesoscopic folds. However, owing to the fact that in many cases the P₁ cleavage is the active surface while the bedding behaves passively, this segregation is not as strongly expressed as in other areas. Another reason perhaps is that the two phases are inclined nearly at right angles, and intersection of trends must occur. Thus, there are many examples of the P₄ cleavage cutting obliquely across the earlier cores producing lineations oblique to fold cores. On the other hand, Plate 22-a is an example of mesoscopic segregation, showing a cluster of P₄ folds in a phyllitic rock intersecting a P₁ fold in a layer of competent arenite. The earlier folds are terminated against the arenite and are not developed in the fold core. The cluster of P₄ folds reappears on the other side of the earlier fold core.

The P₄ structures increase in abundance and intensity westward toward the Keith Metamorphics. Also, the P₄ fold axes and cleavage trend 020° and approximately parallel to similar structures within the Keith Metamorphics. It therefore seems probable that these are of the same generation.

Cooeee Hinge (Figure 28)

The Cooeee Hinge is the zone between the mouth of the Cam River and Cooeee Point, in which the bedding and enveloping surfaces of the P1 folds are very steep. The generalized strike of the bedding, which is clearly seen on the maps, is 078° and faces southeast. The P1 fold axes plunge 10° toward 248° , and there is a slight spread of fold axes in their own axial planes (Figure 28-b). The P1 axial planes have a mean attitude striking north-northwest and dipping 10° west. The P1 cleavage diagram (Figure 28-c) shows a distinct great-circle spread about the fold axis due to the fanning of the cleavage. It is generally inclined at right angles to the bedding.

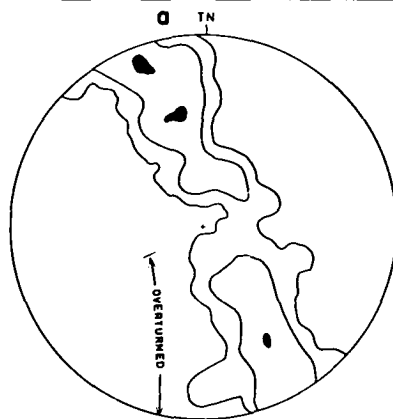
The P1 folds face sub-horizontally to the southeast. The vergence is variable because of secondary reversals on larger folds, but as a whole, the vergence is nil and the folds are generally symmetrical. This zone is the nose of a large recumbent syncline facing south-east, which has the Somerset Overturned Belt as the upper, overturned limb.

The P3 folds make their first appearance in the Cooeee Hinge. These are coupled folds of a concentric style without cleavage. They are disharmonic, often passing into axial-plane thrusts up or down the profile, and dying out along the

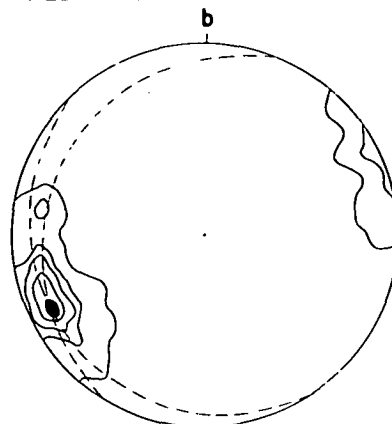
trend by a convergence of the coupled axes. This feature is shown diagrammatically in Figure 32-b. In places a strongly fanning "pseudo axial-plane cleavage" is found in the arenite. This is the refolded P1 cleavage which tends to lie almost perpendicular to the bedding.

The P3 folds are found in the zones of planar vertical bedding, away from the P1 mesoscopic folds, and appear to die out against the earlier folds by a convergence of axes. The fold axes are oriented sub-horizontally to the northeast or southwest. These folds produce narrow zones about 10 feet wide cutting obliquely across the dominant strike, in which the bedding is returned to upward facing. They do not affect the gross orientation of the bedding or the homogeneity of the P1 structures. They have a constant vergence, (Figure 28-e) with an anticline to the southeast followed immediately by a syncline to the northwest, indicating a south-east-side-up movement.

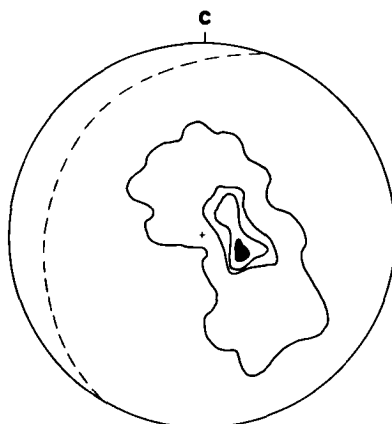
The easterly continuation of the P4 vertical strain-slip cleavage is spasmodically developed in this zone. It has the same orientation as in the Somerset Overturned Belt but is not accompanied by folding. It cuts across the P3 folds without deviation.



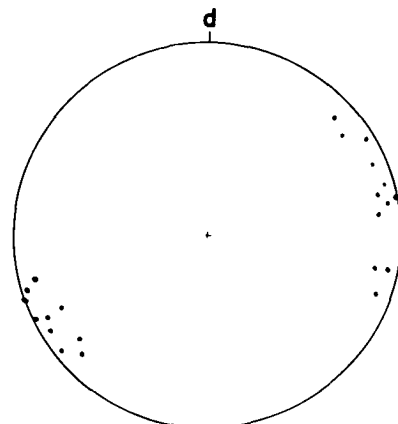
75 poles to bedding.
 $\beta = P_1$ axis = 235/14
 contours 1-3-6%



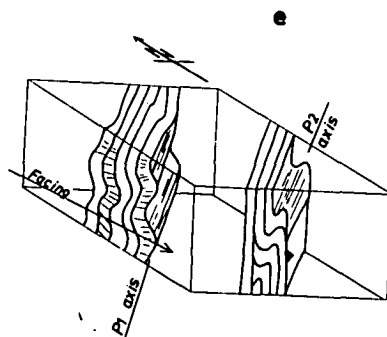
75 poles of P_1 fold axes
 Showing mean range of
 P_1 axial planes.
 contours 1-3-6-12%



75 poles to P_1 cleavage.
 contours 1-3-6-9%



16 poles of P_2 fold axes.



Diagrammatic representation
 of vergence and facing of
 P_1 and P_2 folds.

STRUCTURAL DATA COOEE HINGE

Figure 28.

Parklands Flat Belt (Figure 29)

The Parklands Flat Belt extends from Coeee Point to the limit of the basement outcrop at Burnie. It is characterized by the sudden disappearance of P1 structures, the sudden increase in frequency of the P3 structures, and the presence of intrusive dolerite bodies.

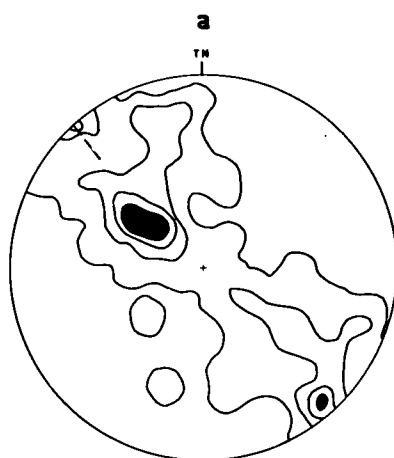
On the western side of Coeee Point, the P1 cleavage is still visible as a coarse anastomosing "fracture" cleavage in the arenite and a weak slaty cleavage in the lutite. At this locality the P1 cleavage is clearly folded by the P3 folds. Only three examples of P1 folds are found in this strip. Consequently, the P1 cleavage is very rare, and except for some minor development of P3 cleavage, the rocks are uncleaved.

Figure 29-a gives the distribution of bedding poles across the Parklands Flat Belt and shows two maxima corresponding to the mean attitude of fold limbs. The general attitude of the bedding, given by the enveloping surface of the P3 folds, is flat lying and upward facing. Rare overturning occurs locally on the steep limbs of the P3 folds. The area is broadly homogeneous with respect to the P3 fold axes which have a northeast-southwest trend with very low plunges (Figure 29-c). The axial planes are generally vertical.

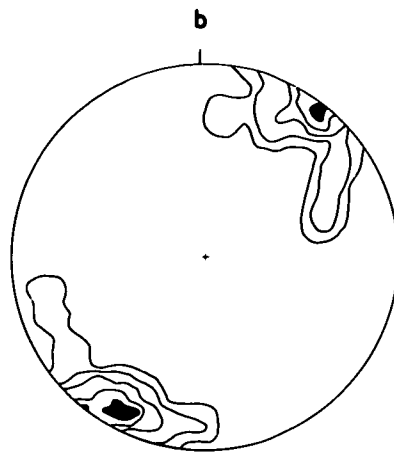
The P3 mesoscopic folds vary in wave length from six inches in the fissile shale to 100 feet in the normal rock type. In style, they are purely concentric and characterized by the general absence of cleavage on the microscopic or mesoscopic scale. Only rarely, is there a fan cracking in the tighter fold cores in the arenite, and a non-penetrative planar cracking in the shale. Rotational joints filled with quartz veins, small saddle reefs and slickensides on bedding surfaces perpendicular to the fold axes are also present.

The plot of fold axes from this belt, (Figure 29-b), shows two maxima, one plunging gently northeast and the other southwest. This is due to an en échelon arrangement manifest by divergent plunges on oppositely coupled folds and also doubly plunging anticlines and synclines. This often produces gentle domes and basins, some of which can be seen on the shore platform at Stoney Creek, and are shown on Figure 55.

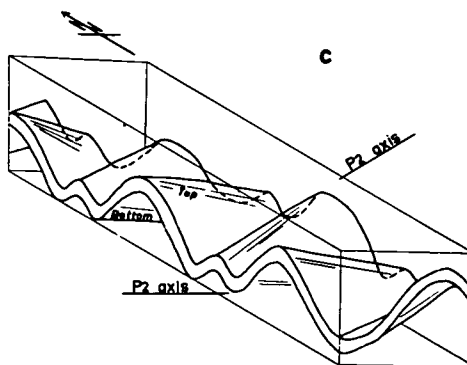
The dolerite bodies are sills or slightly transgressive sheets and it is thought their intrusion preceded the P3 folding. A detailed cross-section is given in Figure 14 and the structural relations of the dolerite have been described in Chapter 7.



121 poles to bedding
contours 1-3-6-9%



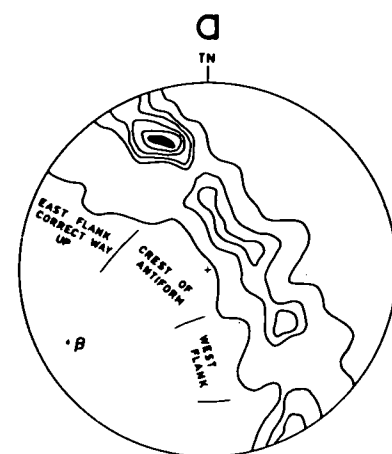
52 poles of P2 fold axes
contours 1-6-9-12-15%



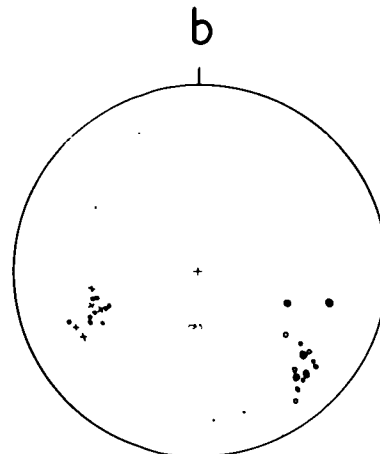
Diagrammatic representation
of P2 folds.

ORIENTATION DATA PARKLANDS FLAT BELT

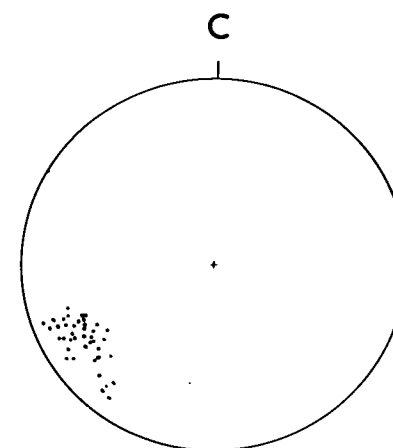
Figure 29.



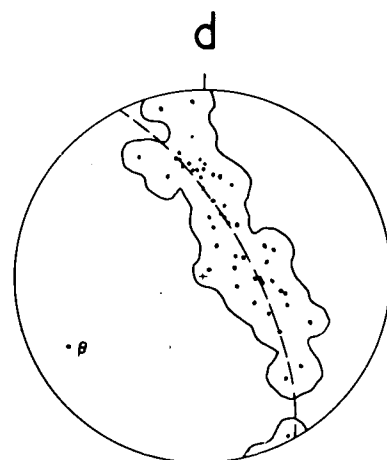
100 poles to bedding
across crest of P3
anticline
 $\beta = 242/18^\circ$
Contours 1-3-6-12-15-20%



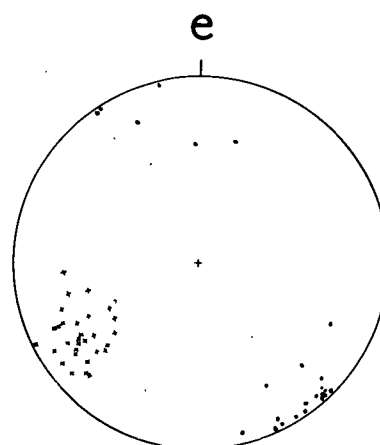
P1 fold axes and axial planes
West flank {
• fold axes
• axial planes
East flank {
• fold axes
• axial planes



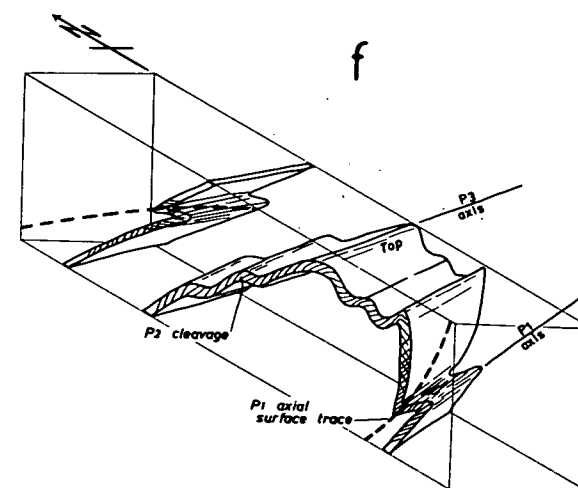
40 P2 lineations
(intersection of bedding and
P2 cleavage)



50 poles to P2 cleavage
disoriented by P3 folding
 $\beta = 243/20^\circ$
showing only 1% contour




29 P3 fold axes
and axial planes
 $\beta = 240/18^\circ$ true





Diagrammatic relationships of
P1, P2 and P3 structures.

STRUCTURAL DATA — ROUND HILL HINGE


Figure 30.



common limbs are flat lying and completely overturned. The vergence is incorrect for them to be genetically related to the anticlinal hinge as shown in Figure 30-f.



Another cluster of P1 folds occurs on the eastern flank, immediately east of the navigation light at Round Hill Point. These are moderately plunging to the southwest, but are upward facing. The outer limbs are steeply dipping, facing southeast, and the common limb is dipping west and upward facing. In Figure 30-b the attitude of P1 fold axes and axial planes on the western flank are compared with those on the eastern flank. There is no significant difference in orientation of these two groups. Hence, a statistical geometrical analysis of both axes and axial planes would fail to detect any subsequent deformation, whereas the vergence and facings show conclusively that the folds have been rotated through 180° about an identical kinematic b-axis during the P3 folding.



The P2 event is a relatively unimportant phase of cleavage development in the interval between the two main phases of folding. It occurs only at Round Hill Point and is not known to be related to any mesoscopic or megascopic folding. The cleavage is confined to the mudstone where it takes the form of either a well-developed crenulation

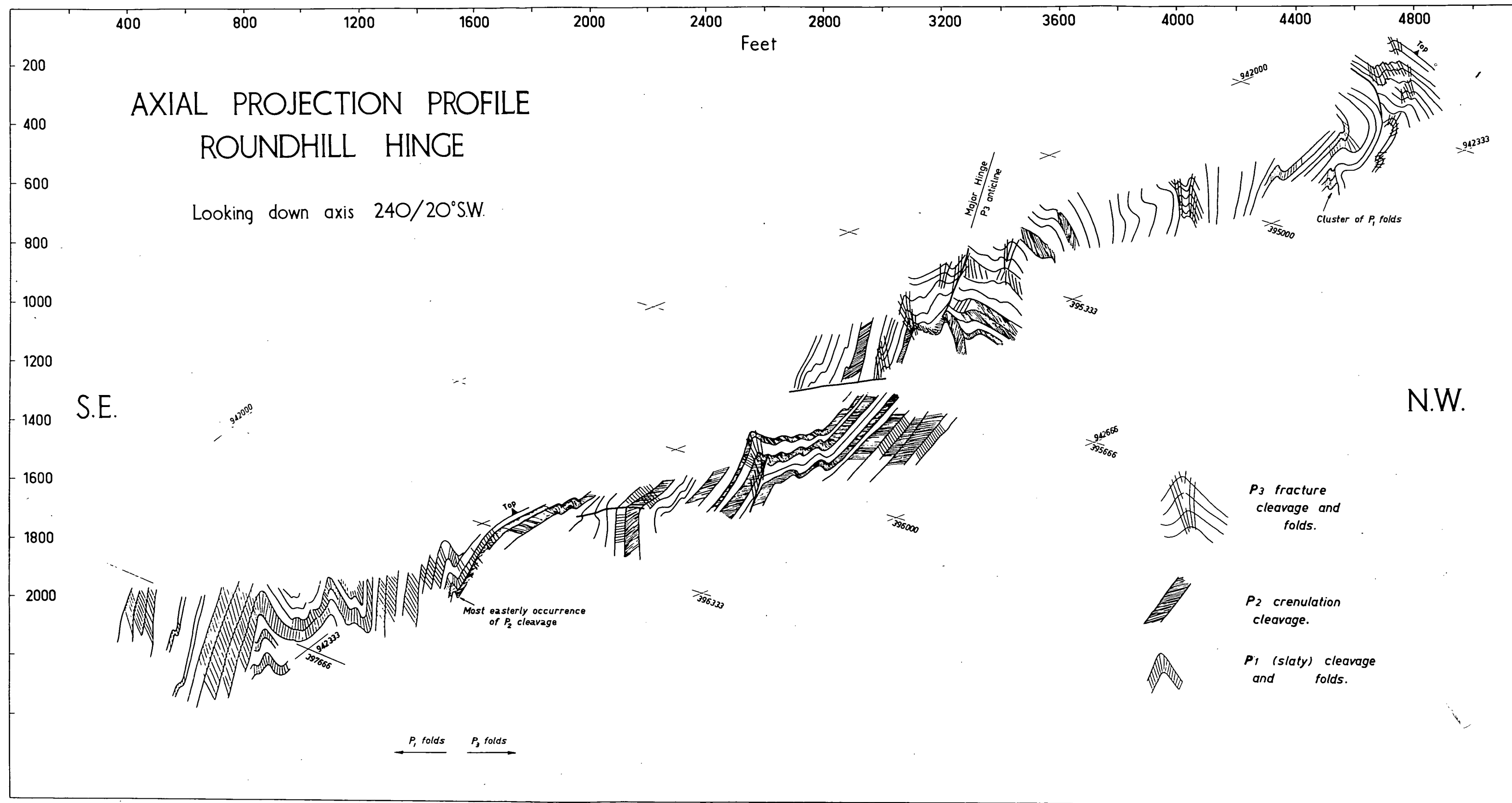


Figure 31.

cleavage if there is a pre-existing P1 cleavage, or a weak slaty cleavage if the P1 cleavage is undeveloped.

The spatial distribution of the P2 cleavage in relation to other structures is shown in the detailed axial projection profile of Round Hill Point (Figure 31). The spatial segregation is so sharp, that on moving along the foreshore platform to the northwest, the first P2 cleavage makes its appearance in the last P1 fold. At this point, (on the tip of Round Hill Point), both the P1 sandstone and slaty cleavage are present, and the P2 cleavage is a closely spaced crenulation cleavage (Plate 12-b). In this specimen (33330), the remnant P1 slaty cleavage is seen as crenulated cross-laminae within the discrete cleavage slices. The actual cleavage planes are zones into which the basal planes of the fine micaceous material has been rotated. The cleavage in the granular bed in Plate 12-b is the P1 sandstone cleavage.

Westward from this point the P1 slaty cleavage fades rapidly, although the P1 sandstone cleavage persists further. Thus, the common occurrence of P2 cleavage in the mudstone and P1 sandstone cleavage in the arenite, both approximately on the same axis, gives the false impression of

extreme refraction of genetically related cleavages.

The P2 cleavage intersects with the bedding to produce a lineation that is very close to the P1 and P3 axes, (Figure 30-c). Accurate measurements reveal a slight difference, for example, in the synclinal fold where the P1 and P2 cleavages interact, the P1 lineation is 245° plunging 50° , and the P2 lineation on the vertical limb is 247° plunging 25° .

Clear examples of folding of the P2 cleavage by the P3 folds are found in the Round Hill Hinge. Thus the poles to the P2 cleavage in Figure 30-d are dispersed in a nearly complete great-circle girdle about an axis corresponding closely to the B-axis for the P3 folding. As a general case, according to Stauffer (1964) this dispersion should define a conical surface if the bedding is kinematically active and the cleavage is passive. However, this example is a specific case where the deformation is coaxial and the cleavage deforms into a cylindroidal surface along with the bedding.

In areas free of P3 folds the P2 cleavage strikes northeast and dips at about 30° . The P2 cleavage has formed as a locally pervasive structure with constant orientation in a zone of planar bedding devoid of pre-existing mesoscopic folds.

The P3 mesoscopic folding is the dominant

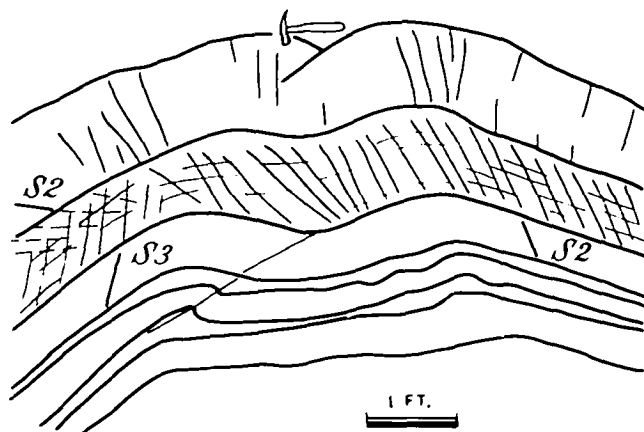
feature of the Round Hill Hinge. All the flat undulating bedding is due to folds of this generation. The fold axes (Figure 30-e) cluster at a maximum of 240° plunging 20° and the axial planes are vertical. The vergence of these folds varies according to whether they are on the eastern or western flank, and are consistent with them being "drag folds" related to the main hinge. The major anticlinal hinge is therefore a third generation structure.

Generally the P3 folds are open and concentric in style, without related cleavage in either the shale or arenite. Commonly they are cylindrical, although many are irregular crumplings of the bedding. A well developed flaggy parting is parallel to bedding and this is probably concentric shearing. Plates 22-b, 23-a and 23-b illustrate the typical style.

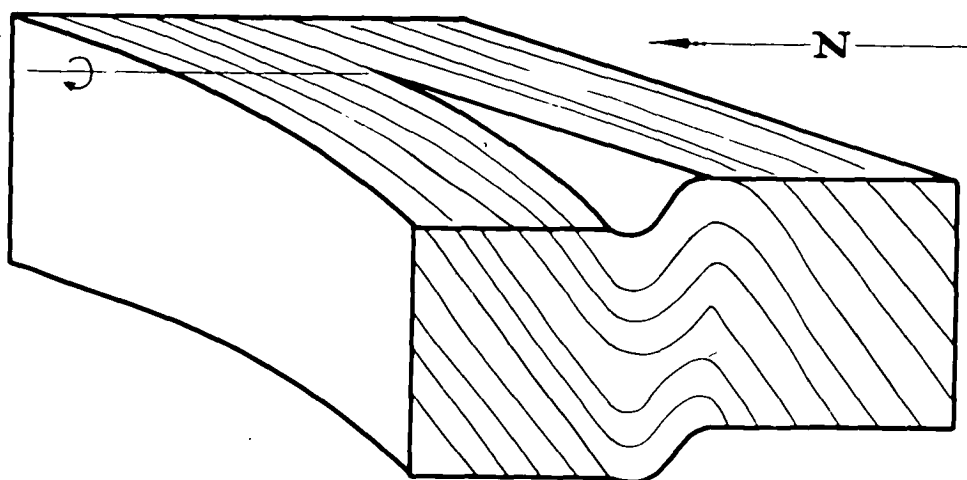
Some of the folds have an axial-plane cleavage, for example, Plate 23-a. This takes the form of a fan cracking in the thicker arenite beds, and a strongly convergent cleavage in the shale. In hand specimen this cleavage in the shale is expressed as discrete smooth planar cracks, whereas in thin section (33328) it is only faintly perceptible as a very weak aggregate polarization of the sericite. the individual sericite flakes are

minute and cloudy. This contrasts with the strong development of P1 slaty cleavage found elsewhere in the same rock type.

An unusual interaction of P2 and P3 cleavages is present in the core of a P3 anticline, at a point 400 yards west of Round Hill Point. At first glance (Plate 23-b, Figure 32-a), it appears that there is a single, strongly convergent axial-plane cleavage. However, the strongly developed cleavage on the right-hand limb, when followed further away from this fold is seen to be itself deformed by folds of similar style and orientation to that with the "pseudo-axial plane" cleavage. This anomaly is resolved by a close inspection of the left-hand limb in Figure 32-a, where an older and almost obliterated cleavage, having the wrong vergence with respect to the fold is mesoscopically visible. In thin section, on the left-hand limb (specimen 33328), the older cleavage is completely destroyed and the new P3 cleavage appears as discrete planar cracks with a weak preferred orientation. On the right-hand limb (33326) there is only the one slaty cleavage expressed by the preferred orientation of mica. Thus it appears that the P2 cleavage is obliterated where it is unfavourably oriented, and accentuated



a. PSEUDO - AXIAL PLANE CONVERGENT CLEAVAGE IN SHALE LAYER IN CREST OF P₃ FOLD AT ROUND HILL POINT.



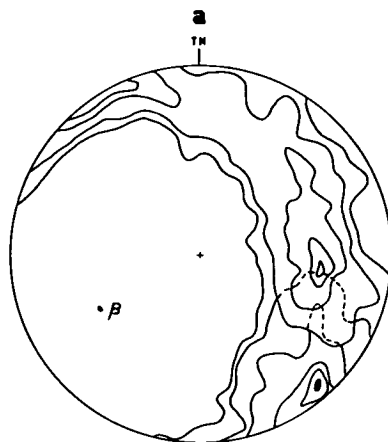
b. DIAGRAMMATIC FIGURE OF COUPLED P₃ FOLD AT FREEZERS POINT SHOWING CONVERGENT ANTIFORM & SYNFORM, & SCISSORS - TYPE MOVEMENT IN FOLD ZONE.

where it is favourably disposed with respect to the P3 fold movements.

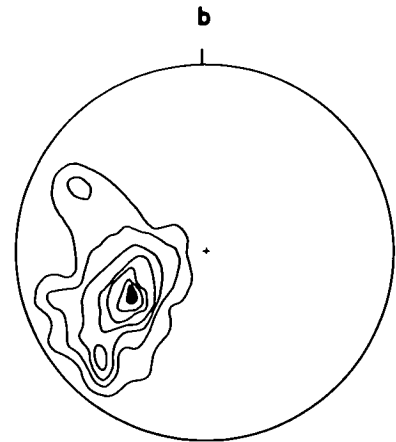
Blythe Overturned Belt (Figure 33)

The Blythe Overturned Belt extends from the navigation light on Round Hill Point, eastward past the Blythe River to the end of the basement outcrop at Sulphur Creek. The belt is characterized by the abundance of P1 folds with moderate to steeply plunging axes, steeply dipping overturned bedding, and the localized appearance of the P5 phase of folding. The greatest thickness of sediment (14,600 feet) is found in this belt.

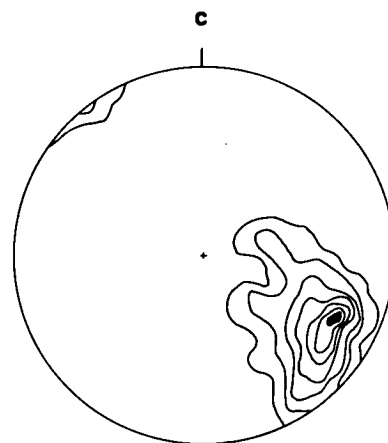
The bedding strikes northeast-southwest, dips steeply to the northwest and is overturned (Figure 33-a). The beds therefore face downward to the southeast. Mesoscopic P1 folds are common, and because of their steep plunge, profiles are best revealed in this area. The axial planes have a strong maximum striking 010° and dipping 60° west, which is more shallow than the bedding. The P1 fold axes have a slight but systematic variation in plunge, being $30-40^{\circ}$ southwest in the western part of the belt, increasing to $60-70^{\circ}$ southwest at Sulphur Creek. There is a strong maximum of fold axes (Figure 33-b) oriented at $240/54^{\circ}$. The pitch angle, measured in the axial plane, likewise varies from 20° up to nearly 90° , so that



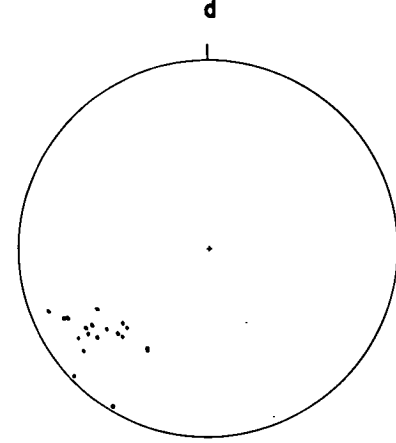
200 poles to bedding
contours 1-3-6-9-12-15%



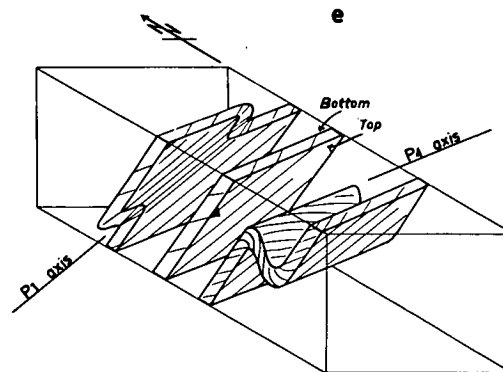
75 poles of P₁ fold axes
1-3-6-9-15-21-27%



65 poles to P₁ axial planes
1-3-6-9-12-15-18%



18 poles of P₄ fold axes.



Diagrammatic representation
of P₁ and P₄ folds.

STRUCTURAL DATA BLYTHE OVERTURNED BELT

they may be described as plunging inclined folds (Turner & Weiss 1963, p.119) with some approaching the reclined position.

The folds are coupled and face upward to the southeast and the enveloping surface dips moderately to the south-southeast. The vergence (Figure 33-c) is invariably constant along the whole strip, except in rare instances when the folds are large, allowing secondary reversals. The vergence, as indicated by the bedding traces in Figure 57, may be described as dextral. The orientation and symmetry of the folds are reflected in the bedding diagram (Figure 33-a). There is a strong maximum representing the overturned straight limbs, a weaker maximum representing the upward facing common limbs, and a complete girdle corresponding to the crests.

The P1 folds generally occur in clusters of coupled mesoscopic folds. An inspection of the detailed map of this zone (Figure 57) shows that the shore platform is divided into zones of folds and zones of planar bedding. The P5 folds tend to be developed in these planar zones.

The P5 folding is of minor regional significance. It occurs only in the area from Blythe Heads to Howth

Railway Station, over a distance of 2 miles. In style they resemble the P3 folds further to the west, however they have the incorrect vergence to be related to the major hinge at Round Hill which is a P3 structure.

The P5 fold axes plunge gently southwest with vertical axial planes. The sense of coupling is such that there is always an antiformal crest immediately to the west of a synformal trough, (Figure 33-c). They are thus truly downward facing with the common limb completely overturned. By virtue of the coupling, the P5 folding does not disturb the overall homogeneity of the P1 structure.

These folds are concentric in style (Plate 24-a), often with a mechanically generated bedding fissility around the curvature and axial-plane break thrusts. Apart from local minor creulations of the P1 sandstone cleavage, they have no associated cleavage. These folds frequently display the same features as the P3 folds, such as the convergence of fold axes in plan, and the opposite plunge of coupled hinges. These folds are therefore indistinguishable in style from the P3 folds further to the west.

Despite the tendency toward spatial segregation of phases, examples of interaction between

P1 and P5 are common. Burns (1964, p.148-149) has demonstrated refolding on the large P5 fold (Burns termed it a P2 phase fold) 400 yards east of Blythe Heads. Burns showed that the scatter of the P1 lineation (intersection of bedding and early cleavage) around the later fold is considerably reduced by flexural unwinding about the later fold axis. On a smaller scale, Plate 24-b shows clearly the refolding of the P1 cleavage.

Figure 34 is a map of the axial traces of P1 and P5 folds on the shore platform between Blythe Heads and Howth. It shows a tendency for the P5 folds to occur in zones of straight planar bedding between clusters of P1 folds. The P5 axial plane traces trend at 040° and thus intersect the P1 traces which trend 020° . Some examples of intersection producing superposed folding on the mesoscopic scale are known from this area, for example, at localities 200 yards east, and 1 mile east of Blythe Heads. The coverage in Figure 34 is linear and not two dimensional, and if this segregation is expansive, it is uncertain how the later folds would die out against the zones of earlier folds. One example (2 miles east of Blythe Heads) is known where a prominent P5 fold trends toward a group of P1 fold cores. At the P1 folds

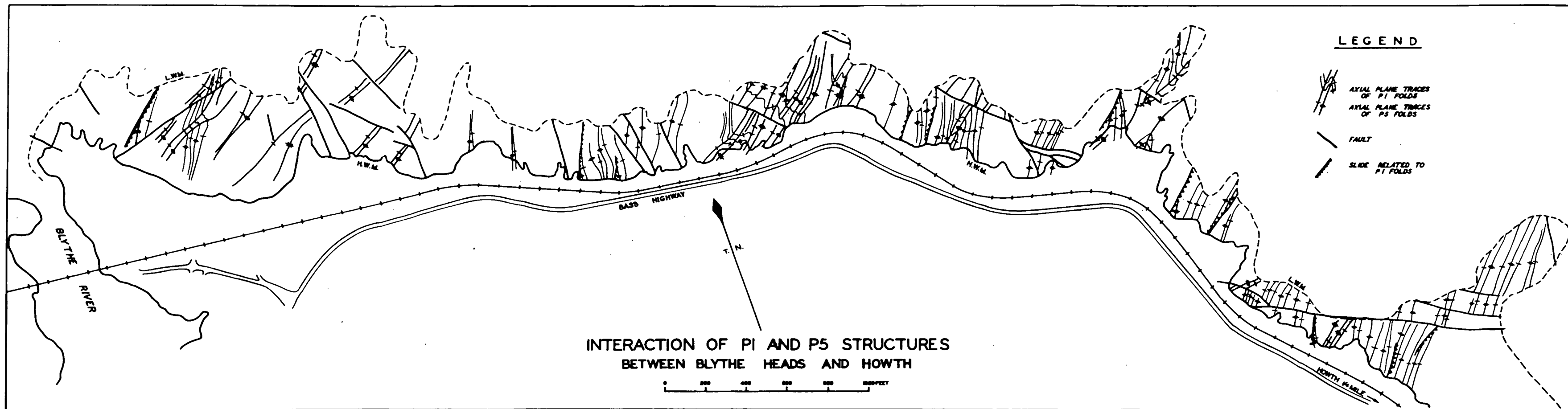


Figure 34.

the later fold vanishes, but is found again in the zone of planar bedding on the other side. The only effect of the later deformation seen in the zone of earlier folds is a system of kink-bands of the P1 cleavage having axial planes parallel to the later fold. It may be that a group of later folds dies out against a group of earlier folds by an en échelon arrangement of sigmoidally curved crests with an overall trend parallel to the elongation of the receptive straight zones.

The relative age of these folds is not clear. They are clearly later than P1 structures and they are not related to the major P3 hinge at Round Hill Point because they have the wrong vergence. They could be earlier or later than the P3 phase but are tentatively assigned as later and hence called P5.

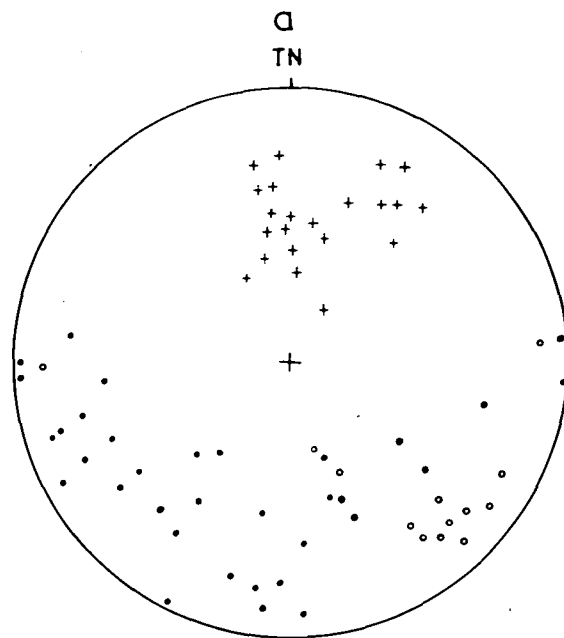
At Sulphur Creek, the Burnie Formation is unconformably overlain by a transgressive sheet of basal Ordovician conglomerate, of the Dial Group (Burns 1964, P.76). The bedding in the conglomerate is gently tilted and in the Precambrian it is vertical. There is no evidence of translation along the unconformity surface and this has lead Burns to suggest that some of the angular style kink-banding in the slate is the equivalent of slight folding in the conglomerate of presumed Tabberabberan age.

Penguin Reversal (Figure 35)

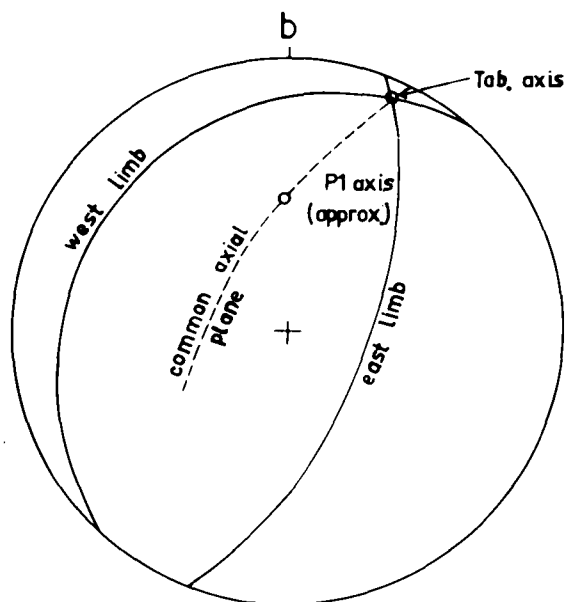
The Penguin Reversal is the least understood of all the structural belts. Between Beecraft Point and Dial Point, the Burnie Formation occupies an inlier in the crest of a Tabberabberan anticline in the Ordovician conglomerate. The Burnie Formation outcrops again beneath the Cambrian just east of Penguin. The Precambrian structure is complicated by the effects of the Tabberabban Orogeny, a major poly-phase movement in the Devonian which is strongly expressed in the nearby Dial Range, (Burns 1964).

This zone is homogeneous in respect to the upward facing of the bedding in contrast to the downward facing of the Blythe Overturned Belt. A major syncline is therefore postulated between Dial Point and Sulphur Creek.

The effects of the Tabberabberan Orogeny can to some extent be removed from the analysis of the Precambrian deformation, because some attitudes of the overlying Ordovician are known. At Dial Point the bedding has a mean strike of 047° and dips 30° northwest, and at Beecraft Point it strikes 020° and dips 66° east, defining an anticline with an axis 024° plunging 8° , (Figure 35-b). There is no minor folding in the conglomerate and there has been no dislocation along the unconformity surface, so that



- 34 poles to bedding
- + 22 P1 fold axes
- o 12 poles to cleavage



RELATION BETWEEN TABBERABBERAN ANTICLINE
AXIS AND P1 ANTICLINE AXIS

STRUCTURAL DATA ACROSS
ANTICLINE CEMETRY HILL
— PENQUIN REVERSAL

Figure 35.

structures in the Precambrian rocks that do not affect the conglomerate must be pre-Ordovician in age. There is a 30° angular discordance at the unconformity on the western limb of the anticline. This contrasts with the 90° pre-Ordovician dip at Sulphur Creek.

Even with the Tabberabberan anticline removed, there is still an anticline present in the Precambrian rocks, outlined by the layer of massive quartzite (inset in Figure 58). Figure 35-a shows the present attitude of bedding across the anticline, giving an ill-defined girdle with a B-axis which has a similar azimuth as the Tabberabberan axis, but a steeper plunge. It appears from Figure 35-b that the P1 anticline and the Tabberabberan anticline share a common axial plane, in which case there should be a tendency toward conical folding, (Ross, 1962).

The minor folds related to the anticline are the earliest folds present here. In the siltstone these folds have a style similar to the P1 folds further to the west, and possess a good slaty cleavage. In the quartzite they are break-thrusted corrugations of the bedding. These folds are assumed to be P1 folds. The regional vergence cannot be deduced because there is a reversal in

vergence of the folds across the anticline. The form of the fold in the bed of massive quartzite suggests a vergence to the east, supporting the postulation of a large syncline between Dial Point and Sulphur Creek.

A second phase of folding is represented by a coupled, near-isoclinal fold with the opposite vergence than the earlier folds. This is situated close to the unconformity on the western limb of the anticline on Dial Point. This clearly refolds the earlier slaty cleavage and has an axis plunging northwest at 50° . There is an axial-plane strain-slip cleavage, quite distinct in style from the slaty cleavage of the earlier folds. This fold cannot be correlated with certainty with any of the fold generations further west and is not assigned to any specific generation.

The eastern-most exposure of the Penguin Reversal is just east of Penguin township where Precambrian rocks occur in contact with Upper Cambrian rocks of the Dial Range Trough (Burns 1964). Burns believed this contact to be an unconformity dipping steeply to the east, against which the "wild-flysch" was deposited (p.51). Here the Precambrian rocks strike northerly and dip 50° - 60° to the east, and are the correct way up.

The overall Precambrian structure in the Penguin Reversal is not clear, but the important point is that the bedding is the correct way up, indicating a synclinal reversal between Sulphur Creek and Penguin, bringing the 15,000 feet of sediment up onto the older Precambrian basement at Ulverstone which is 5 miles further to the east.

It is interesting to note that this tight Precambrian syncline is also the locus of the trace of the Cammena syncline, a Tabberabberan fold in the Ordovician rocks, (Burns 1964, p.76). Thus, in this area there are two examples of control of the Tabberabberan fold structures by the basement structures.

TABBERABBERAN OROGENY

The Tabberabberan Orogeny is a poly-phase movement of Middle Devonian age marking the termination of lower Palaeozoic tectonic movements in Tasmania. Tabberabberan fold structures play a minor role in the evolution of the Rocky Cape Geanticline, and are confined to the marginal zone of the Dial Range Trough which was subject to strong Tabberabberan deformation.

A number of medium-sized folds are defined by the folded angular unconformity at the base of the Ordovician where it rests directly on the Precambrian.

The narrow anticline at Dial Point separates the Penguin Syncline on the east from the Cammena Syncline on the west, (Burns 1964, p.76). These are shown in the section, inset in Figure 58. These all have a northeast-southwest axial trend, parallel to the "grain" of the basement, and the anticline and the Cammena Syncline are spatially controlled by Precambrian structures.

There is no evidence of broad regional warping of the Burnie Formation as a whole that can be attributed to the Tabberabberan Orogeny, as has been found elsewhere in Tasmania by Spry and Gee (1964).

Some minor structures of possible Tabberabberan age are recognised in the Precambrian slate. At Dial Point, an east-west fault which displaces the Ordovician conglomerate is seen to pass into a dispersed zone of kink-folding in the slate. At Sulphur Creek headland, Burns (1964, p.169) described three small fold basins in the Ordovician conglomerate which he correlated with various phases of the Tabberabberan structural sequence.

Kink-folds in the underlying slate appear to be the counterpart of the small fold basins. On the foreshore between Howth and Blythe River are some kink-folds, some of which occur in conjugate pairs. These are spatially related to a set of east-west

faults with a transcurrent component of movement, and are especially strongly developed within fault slices of slate between fault planes. It is not clear whether these are Tabberabberan structures.

MACROSCOPIC SYNTHESIS

For the purpose of deducing the major structure from the detailed mesoscopic mapping, it is necessary to consider only the P1 and P3 structures. The P2 structure, which is a locally pervasive strain-slip cleavage in the Round Hill Hinge, can be neglected because there is no related folding. The P4 structures although locally significant, are only regular undulations of bedding and do not affect the overall attitude of the enveloping surfaces of the P1 folds. Similarly the P5 folding at Blythe Heads does not affect the overall homogeneity of the P1 structure.

If the mean attitudes of the P1 and P3 axes across the entire outcrop belt of the Burnie Formation are compared, it is seen that both plunge at moderate angles in the southwest quadrant. (The same is also true of the P2, P4 and P5 axes). Furthermore, there is also a strong spatial segregation of the individual fold generations, so that there exists regionally homogeneous belts. It is therefore possible to represent much of the

major structure by means of an axial projection profile.

The profile is constructed by projecting all bedding traces onto a plane inclined at right angles to the mean fold axis. This is done by placing a transparent graph paper overlay over the detailed maps of the shore platform and arranging one of the co-ordinate axes parallel to the azimuth of the mean fold axis. Points on the bedding traces are selected and plotted point by point. The horizontal distance between any one of these points and an arbitrary datum line is read directly off the graph paper. The distance below (or above) the datum line this point plots on the constructed projection is given by this distance multiplied by the sine of the angle of plunge. The number of points that need to be projected to define one bedding trace depends on the closeness of folding. For a straight bedding trace, two or three points are sufficient. For a coupled mesoscopic fold it is necessary to project both of the crests, plus two points on the outer limb. This tends to give a diagrammatic representation of the fold, however, it is not intended that this profile should specify precisely the mesoscopic style, but merely demonstrate the profile of the

larger folds that cannot be seen in the single exposure. Every bedding trace that appears on the detailed structural maps , excluding those in the zones of interaction between different fold generations, have been projected on the profile.

The profile has been constructed from Sulphur Creek to Doctors Rocks, with due regard for regional changes in the axial plunge. The profile is presented (Figure 36) as though the observer were deep within the deformed belt and looking up the tectonic axis to the northeast. It is presented this way rather than the usual down-plunge view, in order to observe the convention in this thesis of looking north at sections or plans.

Where there is no axial plunge, as in the Parklands Flat Belt, it is not possible to construct a profile from the data available. In this case a normal cross-section, projected back onto a line perpendicular to the fold axis, has been constructed.

There are two assumptions involved in this type of analysis. Firstly, it is assumed that any closures of bedding on a scale larger than the mesoscopic scale has the same axis as the mesoscopic folds, (Pumpelley's Rule). In some instances, this is justified because of the

correspondence between B-axes and maxima clusters of mesoscopic fold axes. In other instances this is not entirely correct for the major structure is a composite of two deformations. For example, the Round Hill Hinge is a P3 anticline with an axis $229/18^{\circ}\text{SW}$, and the nearby P1 mesoscopic folds trend $234/40^{\circ}\text{SW}$. In this case the P3 axis is used as the projection axis, introducing a slight distortion of the P1 structure. However, the vergence of the P1 folds is still correctly represented.

The second assumption is that the folds are unchanged in profile for large distances along their axes. This assumption is also not completely justified, especially for the P3 folds in the Parklands Flat Belt which are often en echelon, and in the Coosee Hinge where they die out along their axis due to a scissors movement of adjacent planar zones. The axial persistence of the P1 folds cannot be fully deduced because of the moderate to steep plunges, but it is possible to observe a tightening of the fold profile accompanied by a bowing of the fold axis within the axial plane.

Because these two assumptions are not fully met, the profile cannot be considered as accurate and precise as the data from which it was derived. For the purpose of presentation in this

thesis, it is necessary to reduce the composite profile enormously, from a scale of 200 ft. : 1 inch to 2 inches : 1 mile. Thus most of the individual mesoscopic folds cannot be distinguished. Because the detailed structural investigation was a linear traverse and not an areal coverage, it is necessary to extrapolate the bedding traces up and down in order to show adequately the major structure as revealed by the profile. The areas of extrapolation are shown in Figure 36. All these limitations render the profile somewhat diagrammatic but it is still actualistic and shows accurately the manner in which the minor structures fit together to form the major structure.

One valuable property of the profile is that stratigraphic thicknesses project as the true thickness enabling an estimate of the thickness of the Burnie Formation. In the absence of any marker horizon, a structurally homogeneous zone must be selected where all effects of "doubling up" due to folding can be eliminated, thus arriving at a minimum thickness.

The greatest thickness of sediment is found in the Blythe Overtuned Belt where the beds dip very steeply and strike at right angles to the coastline for a distance of 5 miles. This indicates

a great thickness, but much of it is due to repetition by the P1 folding. The thickness can be calculated from the original profile constructions on the scale of 200 ft. : 1 in., by summation of thicknesses of all the straight zones between the clusters of P1 mesoscopic folds. This gives a figure of 12,400 ft. There are, however, gaps in the basement exposure at Howth and Blythe Heads corresponding to an additional 2,600 feet of unfolded sediment assuming an unchanged regional dip. From the profiles of the actual exposure in the Blythe Overturned Belt an overall figure of 15% can be calculated for the increase in thickness due to "doubling-up". Therefore the additional thickness due to gaps in the basement record probably reduces to 2,200 feet giving a minimum thickness for the Burnie Formation of 14,600 feet.

This calculation assumes that all the mesoscopic folds within the outcrop zones between the gaps in the exposure have been observed and corrected for. This is probably valid because the vast majority of folds can be linked in the field with its opposite couple. It is felt that within the zones of outcrop, over 90% of all folds present have been recorded in the detailed maps.

Faulting presents an added complication. There is a set of west-northwest---east-southeast faults with a transcurrent component. The fold structures across most of these faults can be matched, and proper correction can be made. There is also a set of north-northwest---east-southeast faults which cut across the bedding. Matching of the fold structures indicates transcurrent components of variable directions. Many of these faults have very small displacements and are seen to terminate against an arenite bed. The majority displace northeast-side-to-the-northwest in a direction that will give a value of thickness less than the true value. Where correction is not possible by matching of structures, these faults have been ignored. There is another minor set of faults which runs along the bedding in the slate and cuts acutely across the arenite beds. One arenite bed has been triplicated by this faulting, (Plate 34-b). This faulting has only been found in a small zone at Round Hill Point and is not regionally important.

The calculation of thickness does not take into account the flattening component of the P1 deformation, because there is no way of estimating flattening in the zones of planar bedding. This will give a value less than the true thickness.

Finally, it may be stated that the shape of the major structure shown in Figure 36 suggests that the total thickness could be as much as twice the figure given.

STRUCTURAL EVOLUTION

The unusual feature of the structure is the overall geometric simplicity despite the fact that it has undergone five phases of deformation. This is partly because of the nearly coaxial nature of the deformation, and also because the mesoscopic structures of one generation tend to occur in well defined zones, spatially distinct from zones of other generations. This segregation is represented diagrammatically in Figure 37 by solid lines showing the distribution of these genetically different zones. The thickness of the solid line represents qualitatively the frequency of the mesoscopic folds. The broken lines represent the distribution of cleavage when the folds are absent.

The P1 and P3 structures are the only persistent structures across the deformed belt. It is seen that the P3 folds are developed only where there are no P1 structures. The area of overlap of the zones is small or non-existent, in fact, mesoscopic folds of P1 and P3 are not known to interact, and the only superposition is a

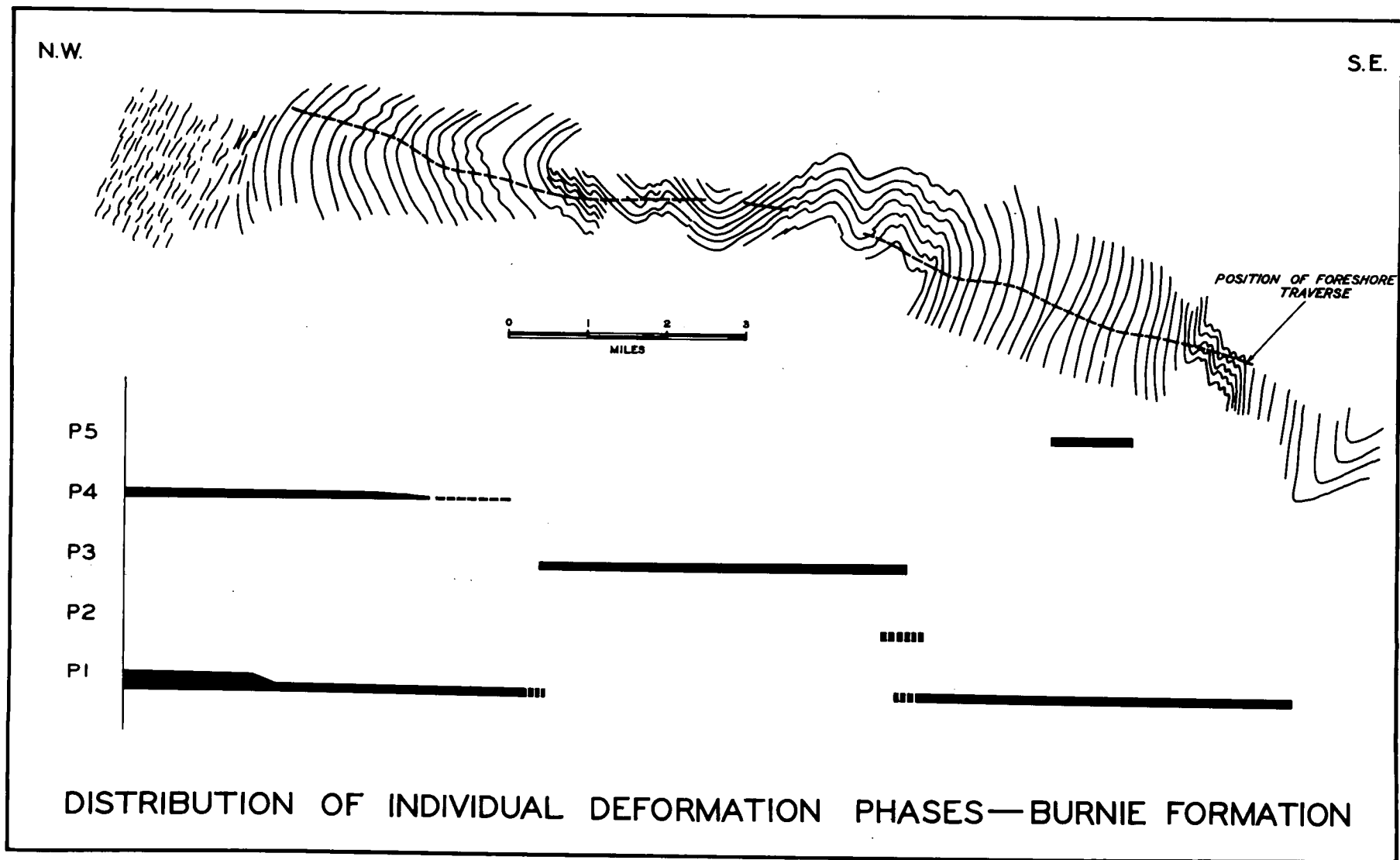


Figure 37.

refolding of P1 cleavage by P3 folds in certain restricted localities at Round Hill Point and Cocee Point. The most spectacular examples of interaction of mesoscopic folds involve the less important later fold generations, for example P1 and P4 interact near Doctors Rocks and P1 and P5 interact at Blythe. The correct interpretation of the structural evolution therefore depends on a few critical localities.

Figure 37 shows a definite correlation between the segregated zones and the major structural elements. The P1 folds are common on the overturned upper limb and in the crest of the recumbent Cocee syncline, but are absent from the lower limb. They appear sporadically in the region between Wivenhoe and Burnie where there is a postulated major P1 anticline. They are virtually absent from the crest of the Round Hill anticline but suddenly become abundant in the overturned eastern limb of the anticline. Mesoscopic P1 folds also occur in the lower limb of the next major syncline to the east at Penguin.

The two main stages in the evolution of the major structure of the Burnie Formation are shown in Figure 38. The P1 primary structure is the result of large-scale tectonic transport to the

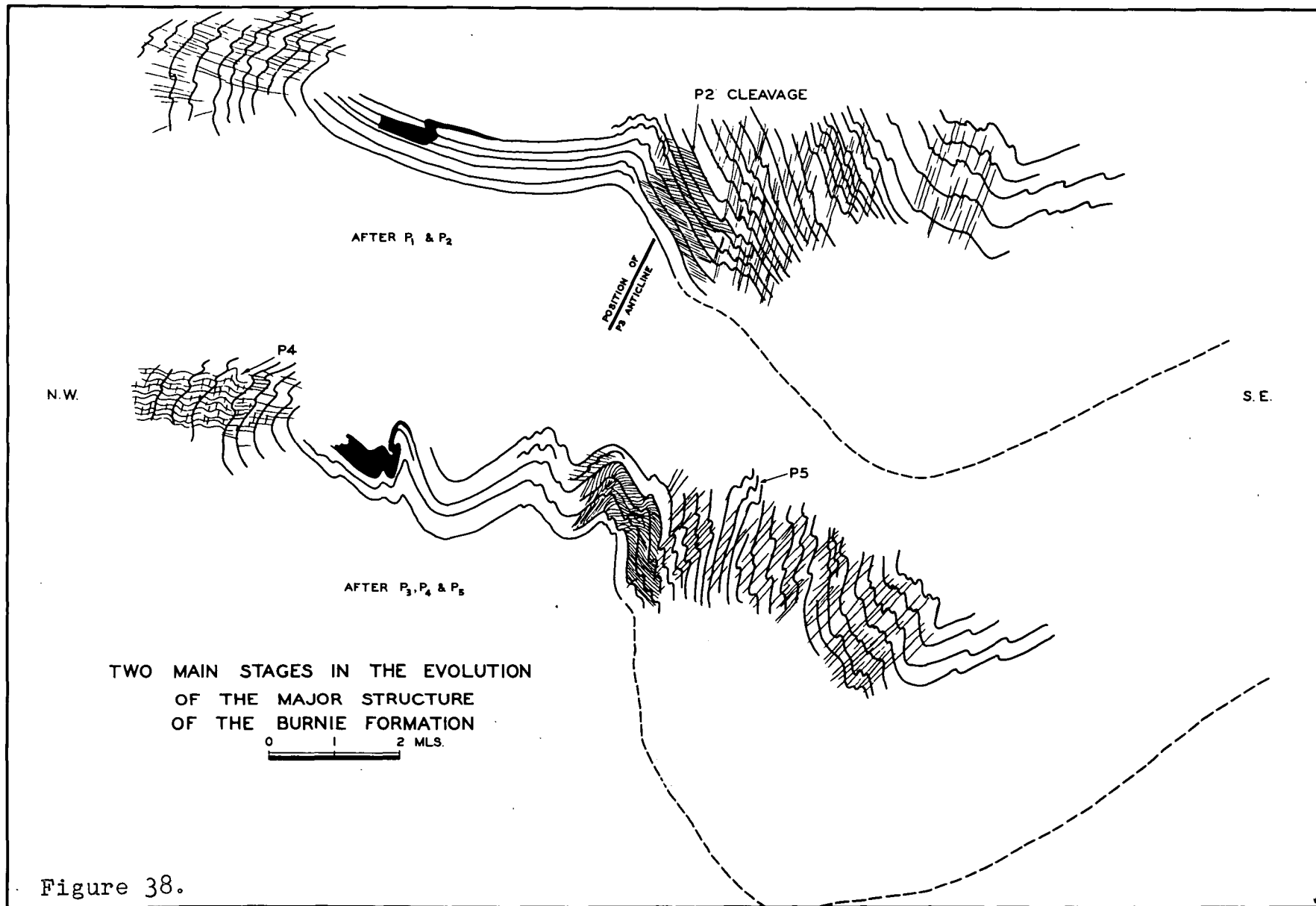


Figure 38.

southeast producing the recumbent Coeee syncline, an asymmetrical anticline somewhere near Wivenhoe, and an asymmetrical syncline somewhere between Sulphur Creek and Penguin. The location of the primary anticline near Wivenhoe is not known exactly. It must lie to the west of the present P3 Round Hill anticline because of the vergence of the P1 mesoscopic folds on the western side of the Round Hill Hinge. Immediately east of the wharves at Burnie are some isolated outcrops, not mapped in detail, containing folds very similar in style to the P1 folds. These have a vergence indicating an anticline to the east. The primary anticline therefore seems to lie between the mouth of the Emu River and Wivenhoe, a two-mile strip of coastline now covered with Recent beach sand.

During this P1 folding the flat belt between the primary hinges behaved passively and remained virtually undeformed. It probably underwent translation and bodily rotation.

The second phase of deformation was merely the locally pervasive strain-slip cleavage or slaty cleavage. It is localized just west of the crest of the primary anticline, where there were no mesoscopic P1 folds.

The third phase of deformation resulted

in continued lateral shortening to the southeast by concentric folding. This deformation was on the same kinematic axis as P1 and P2, and probably the result of the same stress field. The mesoscopic folding was restricted to the flat belt which was free from earlier mesoscopic folding. This is due to the favourable orientation of the bedding. Associated with this lateral shortening was a tightening of the primary anticline by the development of a secondary hinge immediately to the east.

It appears that on reconstructing the pre-P3 structure in Figure 38, a tightening of the syncline forming the Penguin Reversal is required during the P3 deformation. This continued folding of the primary syncline may be reflected by the prolonged sequence of development of minor structures in the upper part of the Blythe Overturned Belt at Sulphur Creek. This sequence, involving a period of cleavage development, minor rotation out of the cleavage plane, disharmonic folding and continued cleavage development is described fully in Chapter 11.

The fourth phase in the evolution was the folding and related strain-slip cleavage development in the extreme western part of the deformed belt. These structures are strongly developed at Doctors Rocks and decrease in intensity to the east and

vanish within three miles. These P₄ structures fringe the Keith Metamorphics which contain structures of similar style and orientation.

The formation of the P₅ folds in the straight zones of the Blythe Overturned Belt completes the structural sequence in the Burnie Formation. They do not fit into the overall picture of continued lateral shortening in a northwest-southeast direction although they are developed about the common B - kinematic axis. It is suggested that they are gravity collapse structures on the steep overturned limb of the fold which locally completely overturned the bedding. It should be emphasised that the position of P₅ in the structural sequence is uncertain; they are clearly later than P₁ and are unrelated to any of the other deformations.

GENERAL CONSIDERATIONS ON SUPERPOSED FOLDING

Comparison of Styles

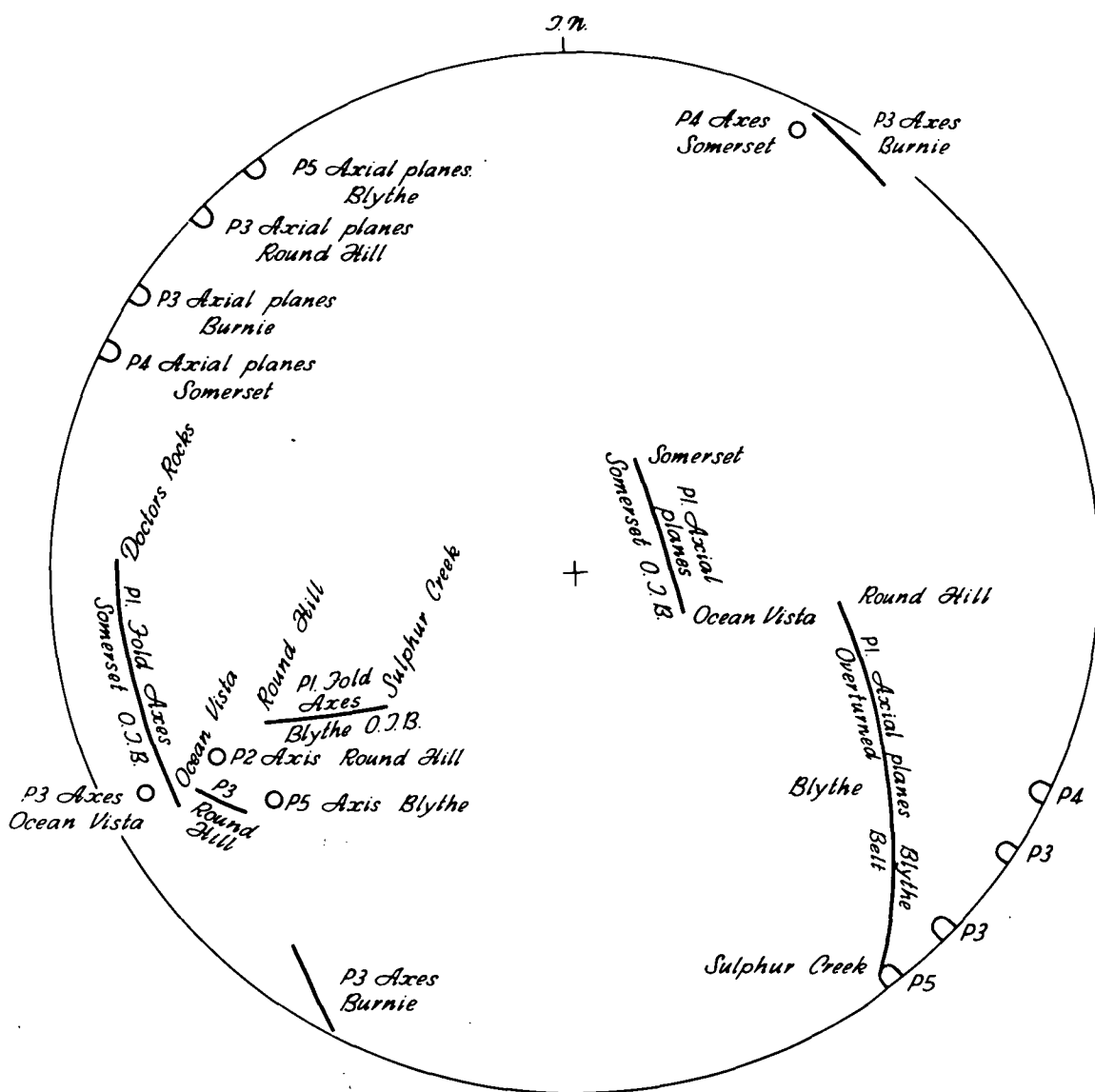
There is a distinct difference in tectonic style of the minor structures between the first generation and the later phases. The first generation is the major tectonic event involving intense penetrative deformation. Minor structures indicate great lateral shortening, initially by buckling, and then by plastic flattening. Widespread, although minor recrystallization of micaceous material occurs.

The later phases are all essentially buckling deformations, with no flattening or recrystallization. The associated cleavages are poorly developed.

This pattern of deformation is frequently found in repeatedly deformed orogenic belts. The characteristic sequence is a first phase of intense similar folding with associated metamorphism, followed by concentric folding without metamorphism, and finally, locally penetrative development of strain-slip cleavages. In cases involving early metamorphism, this behaviour is obviously explained by a convergence of the rheological properties of diverse rock types at higher temperatures, followed by a more elastic response at lower temperatures. In the Burnie Formation, where the metamorphic effects are little more than diagenetic, this early convergence may be due to the unconsolidated nature of the sedimentary pile with the interstitial water playing an important role. Thus, at the end of the P1 deformation the rock fabric probably had condensed to much the present state whereby the quartz grains are tightly bonded by recrystallized matrix and mutually sutured to some extent. During the later deformations, the elastic properties of the rocks would be more significant, and bedding surfaces would be more active, favouring concentric styles.

Comparison of Trends

One of the striking features of all the fold generations in the Burnie Formation is the similarity of axial trends. All axes lie in the southwest quadrant, approximately parallel to the overall trend of the basin of deposition. Figure 39 compares the axial trends of the individual phases, and shows another important difference between the main P1 phase and the later phases. The P1 folds are recumbent, reclined or inclined folds with moderate to steep plunges, whereas the later folds all have near-vertical axes with sub-horizontal plunges. Probably the dominant control over the similarity of axial trends is the result of an overall constancy of the regional stress field throughout the structural history of the Burnie Formation. A lesser control is also suggested by the spatial segregation of minor structures of different generations. It is probable that the strong axial anisotropism generated by the first phase restricted the orientation of the later structures. Taking an extreme hypothetical example of a pile of mullions, it is evident that the most susceptible axis of refolding is around the pre-existing axis.



COMPARISON OF TRENDS IN THE BURNIE FORMATION
POLES TO FOLD AXES AND AXIAL PLANES. ARCS INDICATE CONTINUOUS
INHERENT VARIATION OVER AREA.

Mechanical limitation of repeated folding

Detailed mapping on the foreshore shows a marked tendency toward spatial segregation of the structures of individual generations on all scales. For example, the P3 folds are restricted to the flat belt between oppositely coupled P1 major structures. On smaller scales, there is the confinement of P5 folds to straight zones between clusters of P1 folds, and also the termination of later structures against pre-existing fold cores. The net result is an overall appearance of rapid change of tectonic style, and the general paucity of doubly folded structures.

This in itself indicates the profound change in response to deforming stresses of a sedimentary pile having once been deformed. Although the mechanics of flexural folding, in terms of prediction of precise fold forms or deduction of stress fields, are little understood, it is clear that a planar layer anisotropism is required. During concentric folding approximate constancy of the surface area of the folded layer is maintained, so that the kinematic b-axis lies in the layering. Movement of the kinematic b-axis out of the layering must involve a component of plastic strain in the plane of the layering. Such a situation, whereby the layering then behaves

passively leads to similar type folding (Weiss, 1955, Ramsy^a, 1962), or in some situations, conical folds (Stauffer, 1964).

Considering superposed flexural folding, two types of superposition are possible. Where the later folds are smaller than the initial folds, refolding on the planar segments of limbs is possible, and many examples can be found in the geological literature. Refolding by superposition of larger scale folds is possible, and is expressed as the regional warping of earlier recumbent structures, for example, Cummins and Shackleton (1955), and Spry and Gee (1964). For the other case where both generations are of comparable scales, superposition is impossible for purely flexural folding. This generalization does not apply where the bedding remains passive for the later deformations, so that refolding can occur by similar folding.

It is evident that for compacted sedimentary rocks, the work required for gentle buckling is less than that required by pure flattening because much less extensive granular rearrangements are required. A sedimentary sequence under compression will deform by concentric folding rather than by similar folding if segments of planar layering are available. Thus, according to this concept of

least work, to achieve lateral shortening the later folds will be forced into zones which previously were unfolded.



Plate 21a Photomicrograph of slaty cleavage
in Burnie Formation showing P4
strain-slip cleavage, one mile
west of Somerset. Sp. 33310,
x 27.



Plate 21b P4 dome in Burnie Formation,
one mile west of Somerset.



Plate 22a Group of P4 crenulations (foreground) terminated against P1 fold core in slate (background), one mile west of Somerset.



Plate 22b P3 fold near crest of major P3 anticline, Round Hill Point.



Plate 23a P3 fold near crest of major P3
anticline, Round Hill Point.

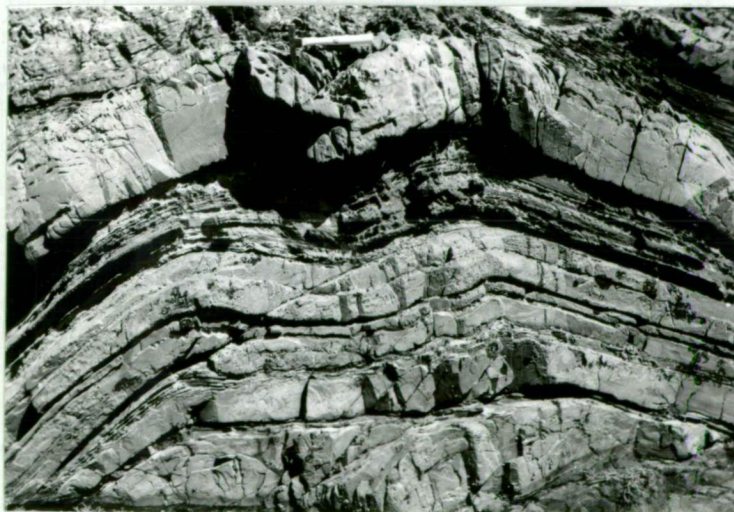


Plate 23b P3 fold near crest of major P3
anticline, Round Hill Point.



Plate 24a P5 fold, between Blythe Heads
and Howth.



Plate 24b P5 fold showing refolding of
earlier P1 cleavage, two miles
west of Howth Railway Station.

CHAPTER 11THE MINOR STRUCTURES OF THE FIRST GENERATION IN THE
BURNIE FORMATIONINTRODUCTION

This chapter is devoted to the description and mechanical interpretation of the minor folds and associated structures of the P1 generation in the Burnie Formation. The P1 phase of deformation was the first and major tectonic event in the evolution of the Rocky Cape Geanticline, during which the Burnie Formation was folded into two large overturned synclines, separated by a flat belt and an anticlinal hinge. The associated minor structures are best developed in the Somerset Overturned Belt and the Blythe Overturned Belt.

The style of the P1 folds, in broad terms, may be described as competent and incompetent folding with marked tendencies toward disharmonic folding. This style is characteristic of flysch-type sediments which consist of a well-bedded alternation of arenite and shale. This type of folding is often loosely described as combined concentric and similar folding, because there is little or no thickening of the arenite layers in the crests, and obvious crestal thickening in the finer grained beds. Such folds possess a good slaty cleavage in the pelite, and a more coarsely spaced sandstone cleavage^(page 279) in the

arenite. These folds display unusual features which give a strong asymmetry on the mesoscopic scale. This is expressed by coupling of folds, common-limb thinning, "half-fan cleavage" and offset carinate structures. Other associated structures include concentric shear joints, oblique shear joints, boudinage and axial-plane faults. The structures are well exposed on the shore platform between Sulphur Creek and Doctors Rocks, and are excellently suited for mechanical studies.

Over most of the outcrop-belt of the Burnie Formation there is a remarkable consistency in size of folds. The majority of folds are coupled with axial separations (orthogonal distance between axial plane varying from 10 feet to 40 feet. Two larger coupled folds, with axial separations of 130 feet and 260 feet, occur midway between Howth and Sulphur Creek (Figure 57). At this locality there are massive arenite layers up to 40 feet thick, interbedded with shale layers up to 200 feet thick, and layers of incompetent pillow lava. These folds have smaller secondary folds with the opposite vergence on the common-limb. The consistency in size of folds therefore seems to be related to the uniformity and regularity of the bedding.

T/57 ^T
FLATTENING OF FOLDS

In the field, all gradations are observed from broad, rounded hinges with no noticeable thickening, to tight hinges showing obvious thickening. In all cases, the shaly beds are thickened in the crests. The folds in profile therefore combine features of both concentric and similar folds. This folding is frequently termed competent and incompetent folding, which by mechanical implication, means that the competent sandstone beds have controlled the fold shapes, and the incompetent siltstone behaved passively, adjusting itself to the interspaces, (Williams, 1961b).

Ramsay (1962) has graphically analysed the thickness changes of small size examples of folded competent and incompetent layers, and compared these with geometrically ideal concentric and similar folds. Ramsay found that some examples fitted closely the pattern expected for concentric folds that have been modified by flattening in a direction perpendicular to the axial plane.

Flattening is the process by which a unit of rock is shortened in one direction by continuous and coherent deformation, accompanied by extension in the plane perpendicular to this direction.

This is defined according to the deformation ellipsoid

in which the maximum and intermediate axes lie in the plane of flattening (A-B plane) and the minimum axis perpendicular to this plane. Campbell (1951), deSitter (1958), Ramsay (1962) and Williams (1965) have applied this to folding, whereby the fold shape is modified by shortening in a direction perpendicular to the axial plane, and extension in some direction within the axial plane.

In the rocks under discussion, it is believed there is evidence of flattening shown by changes of bedding thickness, disharmonic carinate structures and late oblique shear joints. Direct strain indicators such as fossils, oolites and pebbles, are not present, however Ramsay (1962) and Williams (1965) have designed methods, based on deSitter (1958) whereby some quantitative estimate of flattening can be made using properties of the fold profile. Ramsay (1962) considered a pure flexural fold and its flattened derivative (Figure 40-a) and showed that the change in orthogonal thickness, relative to the orthogonal thickness at the apex of the fold, is a direct function of the amount of flattening (Figure 40-b). Ramsay obtained reproducible results using small examples.

The method of Ramsay was tried by Williams (1965) and also by the author, and was found to

have some practical limitations when applied to the mesoscopic scale. This is due primarily to the difficulty of measuring orthogonal thickness (Figure 40-a), which involves the measurement of the distance between parallel tangents. As the points of tangency are seldom directly opposite each other, the errors involved in the direct field measurement can be greater than the variation due to flattening. This difficulty can be overcome to some extent by using an adjustable parallel rule of perspex. In addition, there are observational errors which also tend to be greater than the thickness changes due to flattening. These errors arise because of such factors as erosion on projecting bedding surfaces, graded-bedding, and the presence of joint faces oblique to fold axes. Direct measurement on photographs avoids these errors but introduces errors due to parallax.

Consequently, Williams devised a visual comparison method using a perspex sheet upon which is engraved a semicircle flattened by 10, 20, 30 and 40 percent about the vertical radius. This method has the advantage that it can be used on medium-sized mesoscopic folds, and is capable of detecting arcuate segments indicating differential flattening.

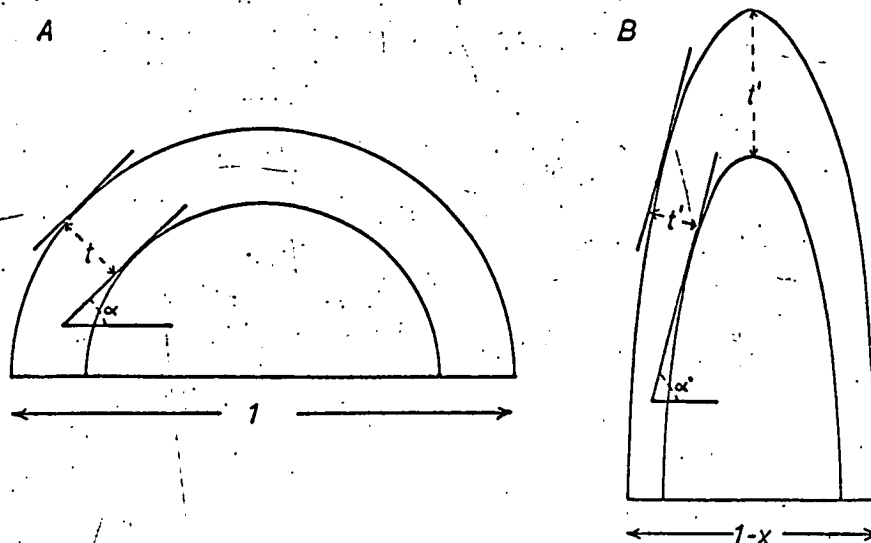


FIG. 5.—A shows a bed of thickness t folded by a concentric flexure; B illustrates how the geometry of this is modified by flattening (x): the angle of dip at any point within the flexure is increased (α to α'), and the thickness of the beds is altered (t to t').

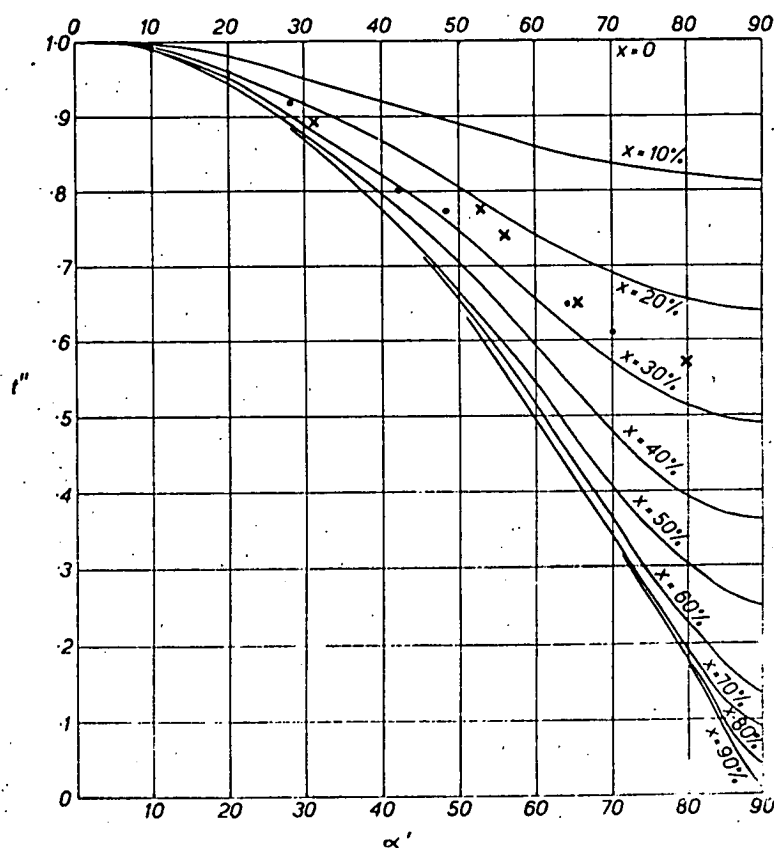


Figure 40. Flattening of concentric arcs, after Ramsay (1962). Points on graph refer to fold in Figure 45. (• left-hand limb, x right-hand limb).

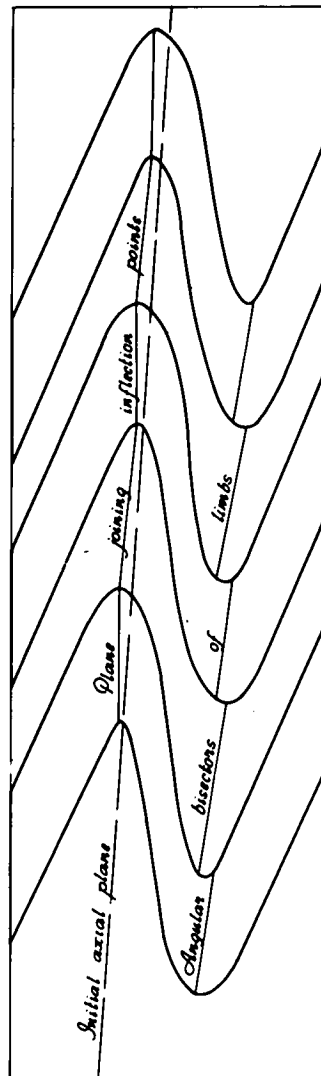
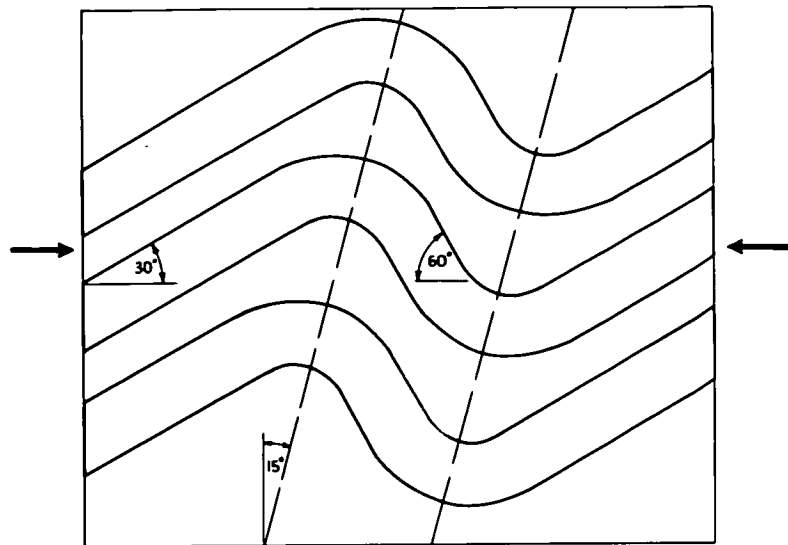
The fundamental assumption of the method of Williams is that the pre-flattened profile is represented by a circular arc. Some writers, for example Rast (1964), believe that the true form of flexural folds is sinusoidal, and this is the basis of the elastico-viscous buckling theories of Biot (1961), and is supported by Currie, Patnode and Trump (1962). However, Williams has shown that some restored fold profiles in the Mathinna Group of northeast Tasmania are in fact flexural folds with circular arcs. The question of the geometrical nature of the pre-flattened profile has not been examined in this thesis. One example of a restored fold profile presented later, approximates a circular arc, but cannot be distinguished with certainty from other types of curves.

Although based on the semicircle, Ramsay's graphical method does not incorporate this assumption. The writer is grateful to D. Leaman, colleague at the Department of Mines, Tasmania, for derivation of the general formula for the graphs of Ramsay (1962) :

$$t'' = \frac{\sqrt{(\tan \alpha')^2 \dots (1-x)^4 + 1}}{\sqrt{(\tan \alpha')^2 + 1}}$$

where the symbols have the same meaning as in

Figure 40. This formula is independent of the initial shape and holds for a concentric fold of any shape.



UNIFORM FLATTENING OF CONCENTRIC FOLDS WHERE
PLANE OF FLATTENING IS OBLIQUE TO INITIAL AXIAL PLANE.

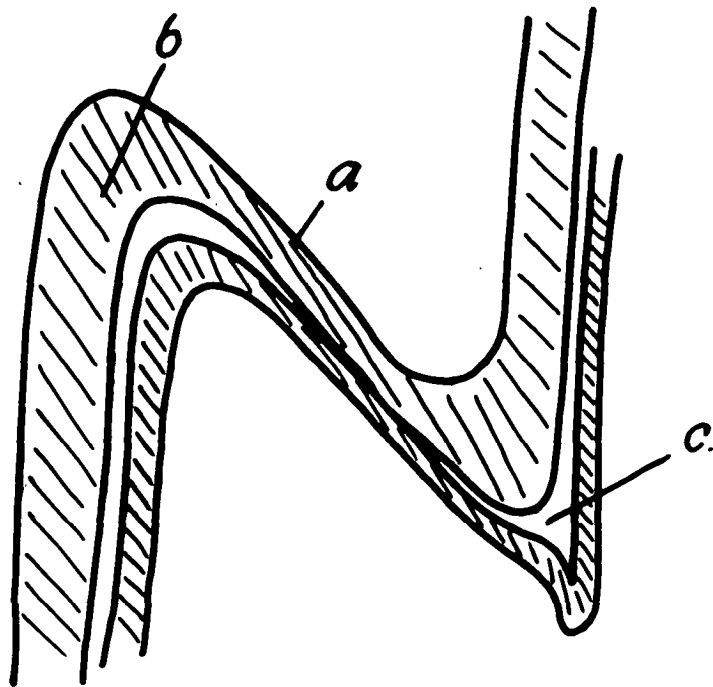
Ramsay (1962, p.315) assumed that the flattening is uniform and that there is no transfer of material along the fold axis during flattening. Unless the initial thickness of the folded bed is known, differential flattening can only be detected but not quantitatively specified by Ramsay's graphs. The assumption must also be made that the layer was planar and devoid of sedimentary lensing.

An important pre-requisite for the application of Ramsay's method is the correct identification of the plane of flattening, because the curves are based on the relationship between layer thickness and inclination of the layer and the axial plane. Ramsay does not define the axial plane and this is difficult for flattened folds where the initial axial plane is inclined to the plane of later flattening. There is evidence, discussed below, to suggest that such a configuration is present in the Burnie Formation.

Figure 41 is a group of hypothetical concentric folds in three competent layers interbedded with incompetent layers. The initial axial plane is the angular bisector of the planar portion of the limbs, and is a plane of orthorhombic symmetry if only the crestal portion is considered. When uniformly flattened in a plane not parallel to the

initial axial plane these folds lose all tendency toward orthorhombic symmetry. The initial axial plane of the folds has no significance in the flattened derivative. The angular bisectors of each of the fold surfaces do not even form a continuous surface. The axial plane is best defined by the continuous but non-planar surface joining all successive inflection points of the fold surfaces. Defined in this way, the axial plane closely approximates the angular bisector in the incompetent material, but in the competent beds, the axial plane and the angular bisector are inclined. In this case the difference is 10° , but for intermediate values of flattening and angular inclinations this may give an error of 50% or even 100% in the value of flattening, calculated by Ramsay's Graphs, (Figure 40).

It can be seen also from Figure 41 that the apical thickness in the competent horizon is the maximum orthogonal thickness and is always contained within the plane of flattening. This follows because the plane of flattening can always be aligned with the radius of the concentric arc. Therefore, providing the plane of flattening can be identified in this way, and providing all the assumptions are valid, then Ramsay's graphical method can still be used. One significant feature



- a. Common-limb thinning.*
- b. Half-fan cleavage.*
- c. Offset carinate structure.*

***SOME FEATURES OF THE
ASYMMETRY OF THE P, COUPLET***

FIG. 42.

of the flattened folds in Figure 41 is that the common limb linking the coupled anticline and syncline shows a differential thinning with respect to the outer limb. This arises because the common limb is initially inclined at a greater angle (60°) to the direction of flattening than the outer limb (30°). This feature is discussed in relation to the folds in the Burnie Formation in the following section.

Asymmetry of the Pl Couplet

On the limbs of the major structures, the minor folds occur as coupled folds with a constant vergence giving an overall monoclinic symmetry to the megascopic fabric. In this chapter the vergence in the Blythe Overturned Belt is referred to as dextral when looking at the outcrop traces on Figures 56 and 57. On the mesoscopic scale the monoclinic symmetry is more strongly expressed by such features as thinning of the common limb of the couplet, an unequal development of cleavage in the arenite, here referred to as "half-fan cleavage", and offset carinate structures. These features are shown diagrammatically in Figure 42.

The differential thinning of the common limb is especially noticeable at Sulphur Creek, but is seen all along the coast-line to Round Hill Point where the vergence is dextral. It is found in both

competent and incompetent horizons on the limb of the two larger dextrally coupled folds at Sulphur Creek, there are secondary reversals showing a sinistral vergence. These folds show a differential thickening of the common limb. In Plate 25-b, measurement of the unit of thinly interbedded siltstone and slate shows that the thickness of the common limb is only 0.7 of the value of the outer limb. The individual laminations within the unit show the same ratio of differential thinning. Consequently, the plane of symmetry (angular bisector of the limbs) is not the true axial plane defined by the surface connecting all inflection points. Precise field measurement of this fold shows that these two surfaces are inclined at seven degrees.

Associated with the differential thinning is the "half-fan cleavage." Plate 25-a is an extreme example. On the outer limb (left-hand limb in the photograph) the sandstone cleavage in the crest is parallel to the axial plane and fans outward on the limb. On the common limb (right-hand limb), the cleavage converges upward toward the axial plane and becomes parallel to the bedding. The amount of fanning varies up to 30° , and the cleavage is not perpendicular to the bedding on the planar portions of the limbs. Thirty-three measurements of the angle between planar bedding and cleavage gave an average

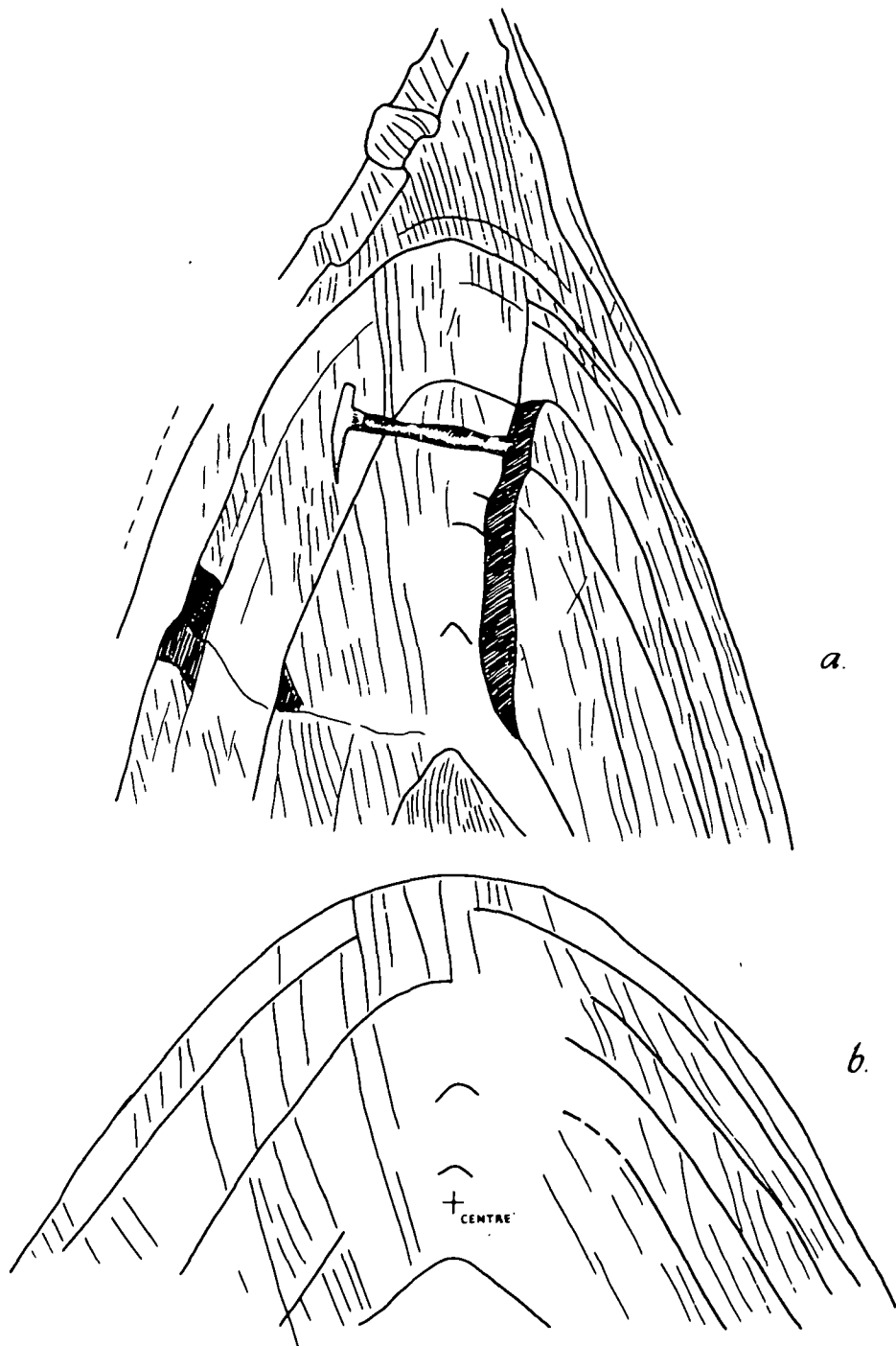
angle of 57° . In thin section the cleavage on the common limb is identical to the normal oblique sandstone cleavage found elsewhere on the fold, and therefore has the same origin as the normal oblique cleavage. It is not concentric cleavage of the type described later in this chapter.

In less extreme cases of common-limb thinning, the cleavage on the common limb is highly acute and contrasts with the larger bedding-cleavage angle on the outer limb. The broad rounded foldcores in thicker competent horizons show an equal and symmetrical cleavage development.

Another feature associated with the differential thinning on the common limb is the non-planar nature of the axial plane. Axial planes of successively higher competent beds appear to be progressively offset in the direction of the common limb, (Figure 41). Also, in Figure 43 the cap of slaty material does not sit exactly on the apex of the fold core in the competent horizon. The cleavage in these slaty cores is oblique to both the angular bisector and the surface joining the inflection points.

Example of flattening in a fold

Several attempts at the estimation of flattening in fold cores in the region between Suplur Creek and Howth were made using Ramsay's method with limited success. The main difficulty is



Anticline at Howth flattened by 25%. (a) Profile drawn from photograph showing asymmetrical nature of cleavage (b) Flattening removed.

FIG. 43.

to obtain accurate and reproducible field measurements of bedding thickness on mesoscopic folds due to the practical limitations discussed previously. Many of the curves when plotted on Ramsay's graph were irregular and frequently entered the "impossible region" of flattening. The example discussed below is one where smooth flattening curves were obtained with good agreement between the graphical and the visual comparison methods.

The fold, illustrated in Plate 26-a, is a plunging anticlinal hinge, situated on the tidal island, 300 yards west of the Howth Railway Station. The fold is coupled immediately to the right by a syncline, so that the left-hand limb in the photograph is the outer limb and the right-hand limb is the common limb. This fold shows many of the features of the P1 folds, such as differential thinning on the common limb, a slight fanning of the cleavage on the outer limb, cleavage parallel to bedding on the common limb, a difference between the angular bisector of the limbs and the axial plane, and an arc of curvature on the common limb that is greater than the arc on the outer limb. Figure 43-a is a direct tracing of an enlargement of the photograph in Plate 26-a, and although no correction for photographic distortion has been attempted, these features are apparent. Parallax error in the

Station	Bedding	α	Thickness	$t'' - t'/\lambda$
1	072/89SE	70 ⁰	5.8 in.	0.61
2	074/86SE	65	6.2	0.65
3	090/77S	48	7.3	0.77
4	094/72S	42	7.6	0.80
5	096/65S	28	8.6	0.92
6	156/53SW	0	9.5	1.0
7	177/43W	32	8.5	0.90
8	202/59W	53	7.4	0.78
9	205/63W	56	7.0	0.74
10	214/65W	65	6.2	0.65
11	225/74NW	80	5.4	0.57

Table 7 : Bed thicknesses for fold shown in
Figures 43 and 45

photograph has accentuated the common-limb thinning.

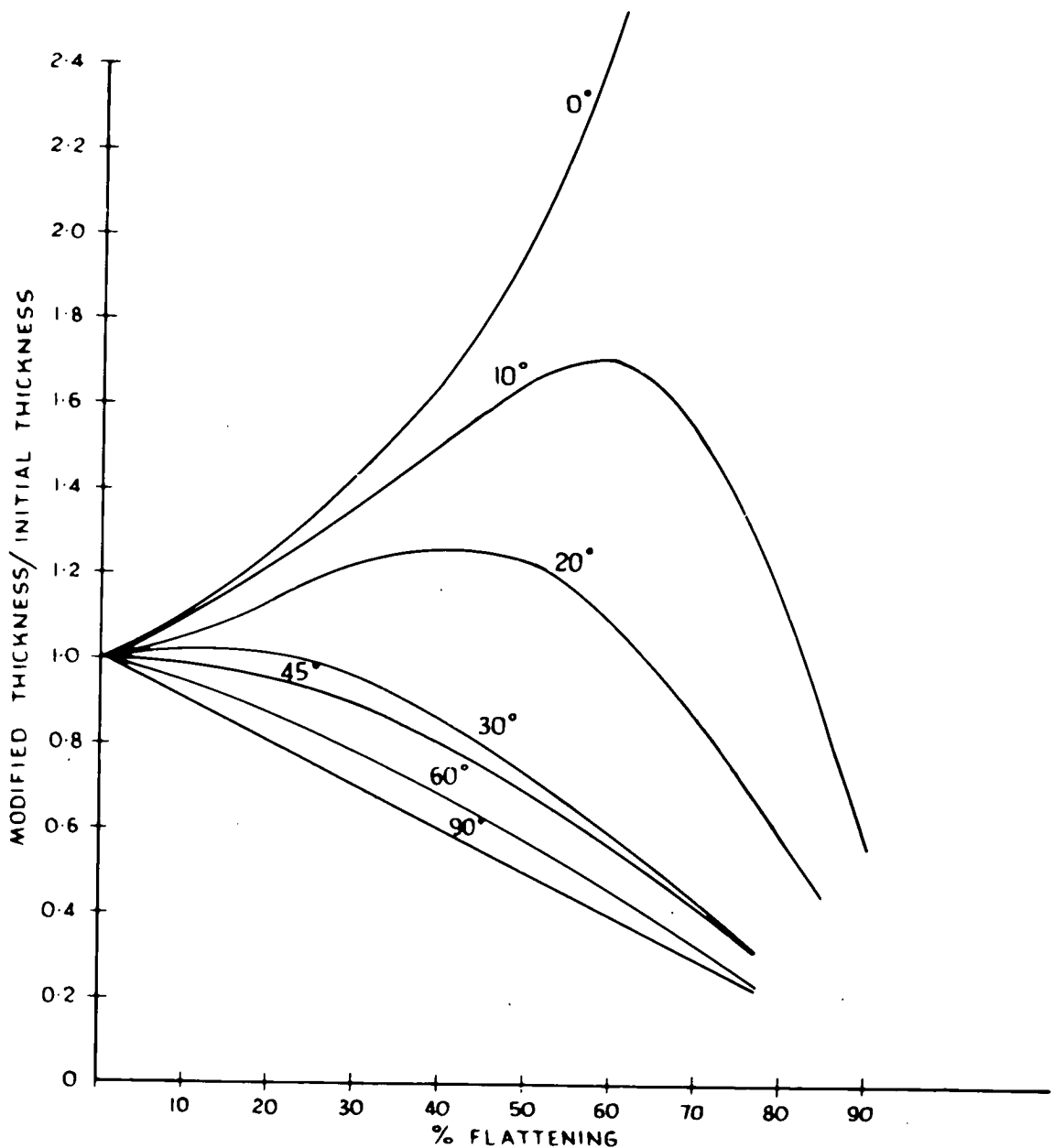
The data for this fold are tabulated in Table 7, where the thicknesses refer to the thin outer parting of the main competent bed. On the common limb the measured unit is obscured by photographic distortion. This fold plunges 53° toward 252° and the axial plane strikes 060° and dips 88° northwest. The angular bisector, deduced stereographically from field measurements, is oblique to the axial plane by six degrees.

Using the visual comparison method of Williams, a value of $25\% \pm 5\%$ is obtained, and no differential flattening is apparent. It is also found that the direct tracing of the top of the competent horizon also fits closely the 25% curve, so that there is no great error due to photographic distortion. The values in Table 7 are plotted in Figure 40-b. The points for both limbs form two curves which approximate to the form of the theoretical curves, indicating uniform flattening. All points except one plot in the broad band between the 20% and 30% curves, in agreement with the value obtained from the alternative method. Deviations from the theoretical curve are attributed to error in measurement.

Interpretation of Mesoscopic Asymmetry

The mesoscopic asymmetry can be explained by the single process of flattening of concentric folds in a direction not exactly perpendicular to the initial axial plane. These features have been reproduced in Figure 41 by flattening a diagrammatic sequence of three concentrically folded competent horizons, interlayered with incompetent horizons, such that the angle between the initial axial plane and the plane of flattening is 15 degrees, measured in the profile plane. In this case, the direction of flattening lies in the profile plane so that the initial axial plane and the plane of flattening intersect in a line parallel to the fold axis. The same general conclusions are still applicable if the direction of flattening lies oblique to the profile plane, but the bedding changes will then record only a component of the flattening. In this case, lineations oblique to the fold axis may result. Evidence of such a movement picture is given later in this chapter.

Differential thinning on opposite limbs of a fold which has undergone uniform flattening is considered to be due to different inclinations to the direction of flattening. Figure 44 is a graph showing the change in thickness of a layer undergoing



Change in thickness of a layer undergoing passive internal rotation (relative to initial thickness) with uniform flattening, for selected values of initial tilt.

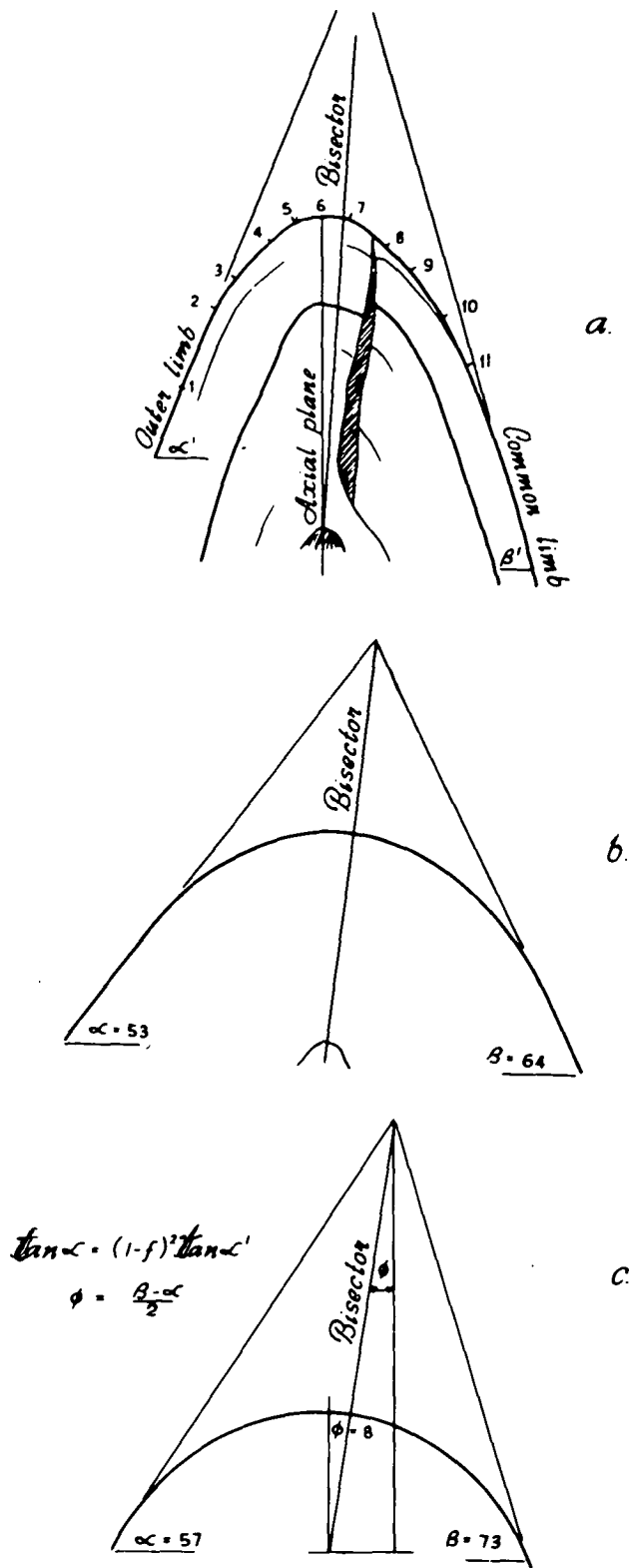
FIG. 44.

passive internal rotation with uniform flattening for selected values of initial tilt. For limbs inclined at less than 45° to the direction of flattening, the thickness will at first increase rapidly up to a maximum, and then decrease rapidly. For inclinations greater than 45° the thickness will immediately decrease, and the beds with higher initial inclinations will decrease more rapidly than those with lower inclinations.

The initial inclinations of the planar segments of the limbs can be evaluated by either of two ways. Firstly, graphic unflattening of the fold profile, restoring it to a concentric fold, gives the initial inclinations with respect to the initial axial plane. The initial axial plane is defined by the angular bisector of the planar tangents. This method requires an accurate representation of the fold profile which is difficult when dealing with mesoscopic folds. Figure 45-b is a restoration of Figure 45-a by graphical removal of 25% flattening. It shows the outer limb inclined at 53° and the common limb inclined at 64° .

Alternatively, the inclinations of both limbs can be reduced to the pre-flattening inclinations by the formula:

$$\tan \alpha = (1-x)^2 \cdot \tan \alpha'$$



a. Tracing of a photograph, b. Graphic unflattening, c. Trigonometrical reconstruction of tangential limbs and calculation of angle between initial axial plane and flattening plane.

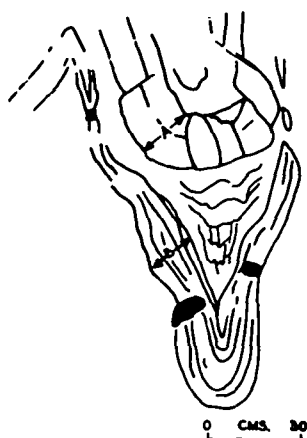
where α is the initial inclination, α' is the final inclination measured in the profile plane, and β is the amount of flattening. The values of α for the planar segments of the limbs are given in Table 7. The pre-flattening inclinations of the planar segments of limbs, with respect to the plane of flattening, are 73° on the common limb, and 57° on the outer limb. The angle (Figure 45-c) between the angular bisector and the plane of flattening is given by $\phi = \frac{1}{2} (\beta - \alpha)$. This gives a value of 8° , compared with 7° obtained by the method of graphic unflattening.

There are alternative explanations of the thinning on the common limb. If the folds are considered purely as flow folds, in the sense of Carey (1954, p.91), in which the directrix (a-kinematic direction) lies within the cleavage plane at an acute angle to the planar outer limb, then the common limb will show a reduced orthogonal thickness while maintaining a constant axial thickness. This explanation is not held because such a movement picture does not account for any features of the fold style, such as regularity of fold size, competent-incompetent relationships, disharmonic profiles, weak development of cleavage on the common limb, and the "half-fan cleavage."

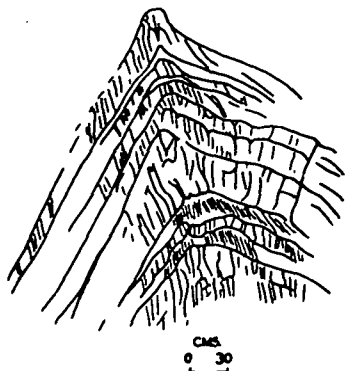
Another explanation that may be advanced requires an overall simple shear which drags the sedimentary pile into a series of asymmetrical drag folds. The common limb would initially lie in the compression direction, but as deformation continues, the common limb may rotate into the tension sector and undergo stretching. This view is not held because in the zones of secondary reversals of vergence on the larger coupled folds, the common limb shows differential thickening with respect to the outer limb. This configuration would not be possible if differential thinning was caused by stretching in the tension sector under the locally prevailing shear couple.

Disharmony

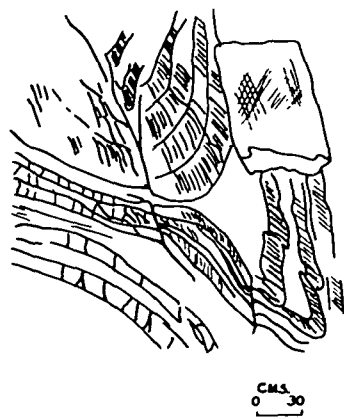
Billings (1942, p.53) used the term "disharmonic" when the fold profile is not repeated more or less uniformly from bed to bed up or down the profile. Disharmony of fold profile is a consequence of the competent and incompetent deformation involving a series of beds which are not of uniform thickness. Disharmonic folding is especially common between Sulphur Creek and Howth where the normal regular layering of mudstone and arenite gives way to a greater variability of bedding thickness.



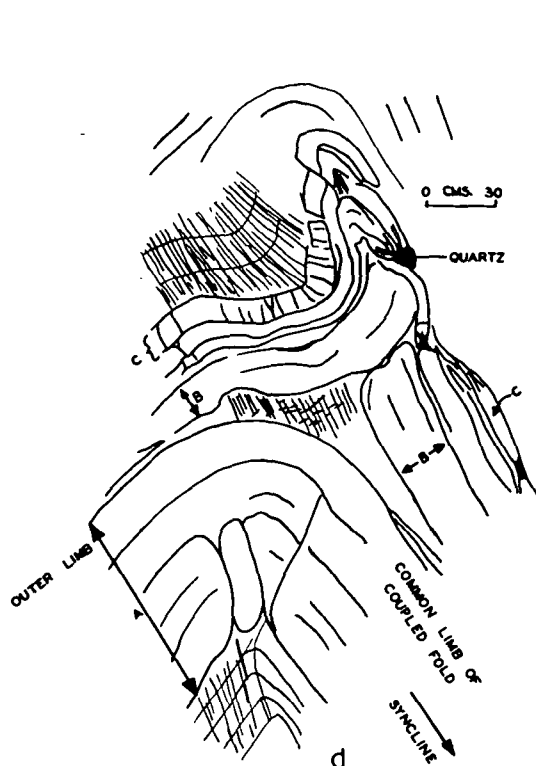
a



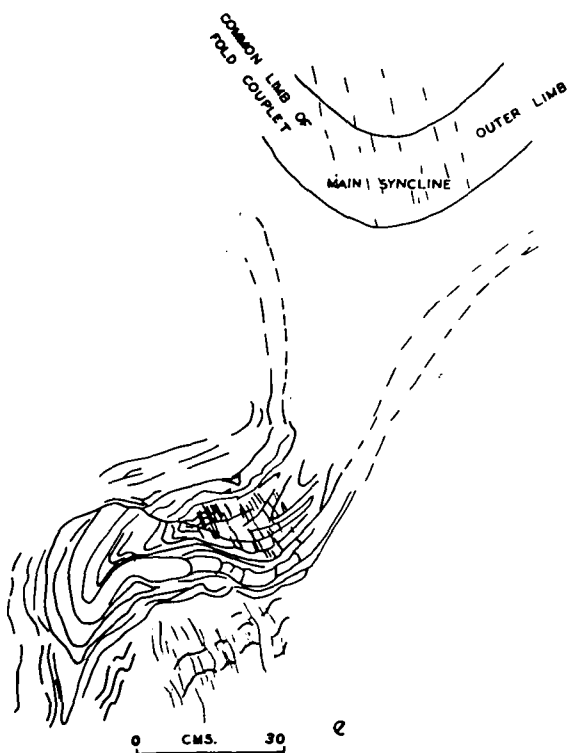
b



c



d



e

DISHARMONIC FOLD PROFILES FROM SULPHUR CREEK

Figure 46.

The tendency toward a disharmonic style is accentuated by two factors, (a) bedding plane slip on one or more surfaces during buckling and flattening, and (b) differential flattening in segments up or down the profile plane. Both these processes result in a structure that may be described as "slip-off", where a thinner competent bed appears to have detached itself from a larger, more rounded fold core in a thicker competent bed. The incompetent bed between is transformed into a wedge-shaped prism which is almost separated from the incompetent material in the limbs. Such structures have been termed carinate cores by Willis (1923, p.20) presumably because of their keel-like shape.

Figure 46-a is an example of the first type of disharmonic structure. The thick competent bed A is folded into a core of larger radius of curvature than the thinner competent bed B, and a carinate syncline is developed in the enclosed incompetent material. Both cores show the same amount of shortening due to flattening (15% according to the visual comparison method), although the core in bed B appears tighter. This implies a greater shortening due to flexural folding in Bed B than in bed A. This must involve movement of bed B away from bed A.

Figure 46-b is an example of the second type of disharmonic structure. The lowermost bed is flattened by 15% according to the visual comparison curve, whereas the succeeding beds shown progressively greater amounts of flattening up to 40%. In this example, the interbedded incompetent beds are thin, and carinate cores are not fully developed.

Figure 46-c is another example of disharmony due to differential flattening. The large synclinal core above is flattened by only 15% but the fold in the thin bed below is flattened by 40%. On the limb of the fold in the thin bed there is a convexity toward the axial plane, a feature to be expected if there is differential flattening up or down the profile. It is noteworthy that this concavity is also reflected in the sandstone cleavage, so that at this spot, the extra 25% flattening below the major core occurred at a stage after the cleavage had formed.

Some extreme examples of disharmonic folding at Sulphur Creek are due to a late stage warping of the "slip-off" structures. In Figure 46-d the amount of flattening in the arenite bed C is the same as in the main competent bed A of the anticline. It is evident that a large amount of bedding-plane slip has occurred. This is especially so on the left-hand limb (outer limb of

fold couplet) of the anticline, between the two arenite beds A and B, where the amount of slip is three feet in a direction along the a-kinematic axis. On the right-hand limb (common limb of fold couplet), the amount of bedding plane slip is much less. Consequently there is disruption of beds B and C, in the form of overthrusting in the core, and boudinage on the common limb. In beds B and C, further down the common limb (not shown in the photograph), there is a regular development of boudinage.

The slaty cleavage in the incompetent horizon above bed C, and also between A and B, is clearly related to the bed in the "slip-off". Evidence of an early slaty cleavage that appears to be axial plane to the deformed carinate structure is found in the incompetent bed between A and B. However the main development of slaty cleavage is congruent with the local trend of slaty cleavage in this area, and is very close to parallelism with the slightly fanned, axial-plane, slaty cleavage in the slaty material below Bed A. The sandstone cleavage in bed C is probably an early feature related to the main folding rather than the late "slip-off."

This "slip-off" structure is still considered to be a P1 structure. The movement picture envisaged, after the initial formation of folds by buckling, is as follows. With continued flattening normal to the axial plane, bedding plane slip occurred along the outer limb toward the crest. The thicker competent bed controlling the initial deformation did not participate in this movement, so the supratenuous beds slipped off forming a secondary buckle. The beds on the common limb remained fixed so that stretching, rupture and boudinage occurred. With one limb moving and the other fixed, a dextral rotationary couple was set up, bending the carinate structure. The outward concavity in the left hand limb of the carinate anticline can be explained by this couple, and was probably assisted by a sealing of the incompetent material between beds A and B, thus maintaining approximately a constant volume while the "slip-off" formed. The sense of rotation of the carinate anticline is the same as that for the coupled fold as a whole.

The final example of a deformed carinate structure is illustrated in Figure 46-c. The sense of rotation is again dextral for both the deformed carinate syncline and the coupled fold.

The slaty cleavage is contemporaneous with the bending of the carinate structure but there is no clear evidence of an earlier cleavage in the axial plane of the carinate fold despite its tightness. This shows that the slaty cleavage of this point formed late in the evolution of the P1 minor structures.

De Sitter, (1958, p.283) considering flattened concentric folds, showed that the total lateral shortening is the sum of the shortening due to concentric folding plus the shortening due to flattening. He states that a bed folded concentrically to the maximum of its capacity 90° dip is shortened 36%, and afterwards shortening takes place by flattening. Actually, this figure of 36% is not the maximum shortening possible by concentric folding since it does not consider bodily rotation of planar segments of limbs. Williams (1965, p.237) considered that flattening proceeds concurrently with flexing because of a collapse in texture in the hinge. Whether these two processes operate concurrently or successively, the final product is the same, and the two components of strain can be treated as kinematic entities.

The examples noted above indicate that the relative proportions of these two components of total shortening can vary from fold to fold, or vary up and down the profile of the same fold. Estimation of the two components can be made at any one spot using either the method of Ramsay or Williams, by first determining the amount of flattening, then graphically restoring the profile and then straightening the folded bed by concentric unwinding.

The evaluation of bulk strain on larger scales would be very difficult, and has not been attempted here. It would require, an extremely accurate plane-table map of one competent horizon over a considerable distance, the amount of flattening for each fold core, and the length and attitude of planar segments of limbs. It then must be assumed that the planar limbs have been flattened by the same amount as the fold crests for it is not possible to determine flattening by these methods, without bed curvature. If granular re-arrangement in the hinge facilitates flattening, (Williams, 1965) then it is possible that flattening will decrease in bands away from the axial plane. Williams (1965) provided some possible examples of this effect within the core region. The reverse may also be possible, and an example is described

later where the limb of a hinge is flattened more than the crest.

CLEAVAGE

The cleavage referred to here as sandstone cleavage appears to be what some writers call "fracture cleavage", however this term is not used because of its genetic implications, and because of the varied use of this term in the geological literature. Leith (1913, p.12) defined fracture cleavage as incipient cemented or welded parallel fractures independent of a parallel arrangement of the mineral constituents. This definition was followed by Nevin (1949, p.163) and many others. On the other hand, de Sitter (1956, p.98, ;.273) and ⁿmay others, believe fracture cleavage passes into slaty cleavage, thus admitting to a certain degree of mineral orientation in fracture cleavage. All writers however, emphasise the discrete nature of the cleavage planes. A mineral orientation is an essential part of the sandstone cleavage under discussion and is believed to differ from slaty cleavage only in the degree of penetration of the rock fabric. This is controlled by grain size and composition. Consequently the term sandstone cleavage is used here instead of fracture cleavage.

Mesosopic aspect of sandstone cleavage

There are several different morphological aspects of the sandstone cleavage in the Burnie Formation. It may be expressed as a planar cleavage parallel to the axial plane and developed equally on both limbs of a fold. This type is transitional into a slight fan cleavage which diverges upward away from the anticlinal core. Another aspect, which is very common in the Blythe Overturned Belt is the "half-fan cleavage" mentioned previously. A less important aspect is a concentric cleavage parallel to the lithological layering.

In the outcrop, the sandstone cleavage appears as closely spaced discrete planes of parting, 1-10 mm apart, along which the rock splits when struck with a hammer. The planes of parting are rough and undulating, due to an anastomosing character which actually splits the rock into thin discoid-shaped slices. The anastomosing nature is seen in both profile plane (a - c fabric plane) and on the bedding plane (b-c fabric plane).

The sandstone cleavage is bedding-contained. At sandstone-siltstone contacts the orientation changes abruptly, becoming more acute to bedding in the slate. This is the "refraction" of some writers. Commonly within

the arenite bed, the apparently discrete cleavage planes are not continuous but are terminated against concentric shear planes. Also, the cleavage is commonly deformed by movement on these concentric shear planes giving a sigmoidal shape in profile, for example, Plate 25-a.

The fold in Figure 43 (a direct tracing of Plate 26-a) is taken as a typical example of a "half-fan cleavage." The cleavage on the outer (left-hand) limb fans at up to 5° to the axial plane. On the common (right-hand) limb, the cleavage is approximately planar, and as the bedding swings from the crest down to the limb, the cleavage changes from highly acute, to parallel to bedding. The cleavage therefore approximates to the simple case of being parallel to the axial plane, or the plane of flattening.

Figure 43-b shows the profile of the same fold with the 25% flattening graphically removed. This procedure is not considered precise because of parallax errors in the photograph from which the figure was traced. However, the gross aspect of the bedding-cleavage relationship is preserved. On the outer limb the cleavage fan is accentuated, although in the planar segment of the limb the amount of internal rotation during unflattening is not nearly sufficient to bring the cleavage

into a perpendicular attitude to the bedding. On the common limb where the cleavage is slightly convergent upward, the acute angle between bedding and cleavage is enlarged, forming a greater degree of upward convergence. Thus, in the restored profile, the relatively simple geometrical relationship between fold form and cleavage, in which the cleavage is approximately parallel to the axial plane, is lost. The centre shown in Figure 43-b is the centre of curvature of the best circular arc than can be drawn on the restored profile. The fact that the centre falls within the arenite bed suggests that the pre-flattening arc was not precisely circular.

This contrasts with the findings of Williams (1965) who found that for some folds in the Silurian rocks of northeast Tasmania, the cleavage restored to an attitude perpendicular to the outer arc. From this it was argued the sandstone cleavage developed during the flexural stage of folding and not during flattening. The configuration of the cleavage in Figure 43-b is an argument in favour of a theory of cleavage development during flattening. In this case, the sandstone cleavage formed parallel to the plane of flattening, or perpendicular to the direction of flattening.

It appears that some deformation by continued concentric folding proceeded concurrently with flattening. During this process the mechanically generated concentric shear planes continued to be active, producing the bed-delimited nature, and the sigmoidal nature of the sandstone cleavage, (page 280/281). The fanning of the sandstone cleavage may then be due to a continued tightening of the fold by bodily rotation of the limb, and perhaps accompanied by differential flattening in the core region.

Microscopic aspect of sandstone cleavage.

Although in hand specimen the cleavage appears as discrete cracks, in thin section it is seen to penetrate on a scale down to the individual quartz grains. The cleavage is primarily due to discontinuous, anastomosing, shred-like aggregates of the fine micaceous matrix material. A preferred orientation of minute crystallites of sericite and a preferred orientation of detrital muscovite also contribute to the fabric. Plate 27-a is typical of the sandstone cleavage in thin section over most of the outcrop belt of the Burnie Formation. A more pronounced development of cleavage, bordering on a phyllitic or schistose foliation, is found in the Somerset Overturned Belt adjacent

to the Keith Metamorphics. This is described later.

The shreds which define the microscopic cleavage planes are actual aggregates of barely resolvable micaceous material, and are not fractures in the usual sense. These aggregates show a brown iron oxide staining which commonly masks the interference colours of the mica. The shreds bifurcate at the detrital quartz grains and enclose lozenge-shaped slices of matrix and granular material of a thickness determined by the largest quartz grains. The angle of bifurcation varies from 5° to 30° . The anastomosing nature is therefore a direct result of the textural features of the sedimentary rock.

Plates 27-a and 27-b respectively are sections cut parallel and perpendicular to the fold axis, and both in a plane perpendicular to the cleavage. This similarity indicates that the lenses enclosed by the anastomosing shreds are discoid slices. Thus, the microfabric has a tendency toward a statistical radial symmetry about an axis normal to the cleavage, and not a monoclinic or orthorhombic symmetry with a symmetry plane normal to the fold axis, as in types of conjugate cleavage, described by Knill (1960, p.320), Rickard (1961, p.330), and Turner and Weiss (1963, p.466, p.493).

The individual micas within these shreds either have an apparent random orientation, or a weak preferred orientation with a direction of aggregate polarization slightly acute to the trend of the cleavage shred. In detail, (Plate 28-a), the bifurc^ation is due to an en échelon arrangement of individual micas which maintain an orientation subparallel to the mean trace of the cleavage. This observation is important because it shows that divergent cleavage shreds were developed synchronously, and are not the result of a process whereby the initial cleavage plane is rotated out of its plane of formation during progressive strain and cut by successively formed planes.

Behaviour of quartz grains:

Petrographic evidence indicates that the quartz grains underwent little change during cleavage formation and behave merely as rigid passive bodies in a flowing matrix. In the coarser quartz wacke from Sulphur Creek, the original size and shape of the detrital quartz can be recognised by the presence of secondary overgrowths (Plate 9-a). These grains have corroded margins due to encroachment of the matrix. The shape of these grains, where no overgrowths are preserved, is still commonly suggestive of rounded grains. There is no evidence

of border granulation, and both undulatory extinction and fracturing are at a minimum. It is evident that point-to-point contact between grains was negligible, and most deformation occurred in the matrix.

Even in the quartz wacke of fine- to medium-sand size, where some point-to-point contacts and suturing are present, the quartz grains do not appear to have been greatly affected by deformation. These arenites still contain up to 30% matrix. The quartz grains are generally equidimensional with rough, fuzzy borders due to matrix corrosion. Some grains show fracturing (34018), although there is no evidence of disruption into sub-grains and no growth of recrystallized matrix in the cracks.

Quartz grains showing deformation lamellae and undulose extinction are scattered sporadically throughout these rocks, but are unrelated to any conspicuous zones of stronger deformation. It is suggested that some of the quartz grains in the Burnie Formation are derived from the orthoquartzite formations to the west, which contain grains showing intragranular strain effects. It seems likely that the strain effects in the clastic grains of the Burnie Formation are inherited features and not related to cleavage development.

A dimensional orientation of quartz grains is not obvious since most grains are equidimensional. However, there is a weak tendency for the more inequant grains to be oriented with their longest dimension in the cleavage plane. This tendency is seen in sections both parallel to, and perpendicular to the fold axis (Plates 28-a and 28-b). This is probably a mechanical rotation.

Behaviour of detrital mica :

Detrital muscovite generally occurs as thin flakes which are generally 10 or even 50 times larger than the average size of the quartz grains. The detrital nature of these flakes is shown by the pinching between quartz grains, fracturing and splintering, and their close association with fine, barely recrystallized sericite matrix. A preferred orientation of muscovite flakes parallel to the cleavage is common, and is especially evident where the cleavage is at a high angle to bedding. This preferred orientation is considered to be a purely mechanical rotation of rigid flakes. Many show slight effects of this rotation such as bending and splintering.

It is of interest to mention an example of preferred orientation of muscovite, taken slightly out of context, from the correlate of the Cowrie

Siltstone in the core of the Sisters Hills anticline in the Rocky Cape Group. The detrital muscovite varies up to 2 mm in length, in marked contrast to the average grain size of 0.1-0.05mm in the siltstone, (Plate 20-a, 33254). In this case the cleavage is at a high angle to the bedding, yet the micas show only slight splintering and bending. This raises the problem of how these large clastic micas rotated through large angles in a granular material and yet remained virtually unaffected. This suggests that the siltstone must have possessed great mobility at the time of deformation, with intergranular bonding and friction at a minimum.

Muscovite in the arenite of the Burnie Formation also occurs in stacks of plates with a tabular or equidimensional outline. These generally lie with the basal plane in, or close to, the bedding; and thus when the cleavage is at a high angle to bedding the muscovite books have a cross-fibre attitude. These cross-fibre books behave as porphyroclasts, being wrapped by the anastomosing cleavage planes. Commonly, (Plate 29-b) the mica is kinked, indicating lateral shortening in a direction perpendicular to the cleavage plane. This shows that flattening was

operative during cleavage development.

Role of matrix :

The matrix, which consists of a pasty mass of sericite, chlorite, fine quartz chips and irregular patches of barely resolvable semi-opaque material, has accommodated most of the strain during deformation. The matrix material shows varying degrees of recrystallization. Within the lensoid cleavage slices, the individual flakes are dirty and barely resolvable, with a weak or no preferred orientation, and generally do not show aggregate polarization. Along the fringes of the quartz grains, especially in the shreds pinched between grains, the individual micas appear as clear crystallites with a strong planar orientation sub-parallel to the cleavage. In the pressure shadows of the quartz grains, clear crystallites of sericite are oriented in the cleavage direction and penetrate slightly into the quartz. This is the clearest expression of recrystallization in the arenite and indicates that recrystallization played some part in the development of cleavage.

Behaviour of clay aggregates and mudstone inclusions:

The rounded aggregates of clay mineral, of about the size as detrital quartz, commonly appear unmodified, showing no sign of shearing or flattening. Thus, these particles have behaved in a manner similar to the quartz grains. This is unusual and may be an indication of low viscosity of the arenite during deformation with only minor intergranular bonding and friction.

The behaviour of the larger mudstone inclusions is different. These inclusions are believed to be fragments torn from the underlying mudstone and incorporated into the arenite by turbidity currents during deposition. The larger inclusions possess a good mesoscopically visible slaty cleavage congruent with the sandstone cleavage in the arenite.

Specimen 33319 is an arenite with a well-developed cleavage perpendicular to bedding, and containing black mudstone inclusions up to 6 mm in length. There is a strong tendency for these inclusions to be elongate in the cleavage direction (Figure 29-a). The average elongation seen in the thin section is 4 : 1 in the cleavage plane. The discrete cleavage planes in the arenite pass straight through the mudstone inclusions, and display a

"bow-tie" structure due to a pinching together of the cleavage planes within the inclusion. These features are indicative of flattening on the microscopic scale, although the "bow-tie" structure suggests that the flattening on this scale was not homogeneous, and was probably greater in the vicinity of these mudstone inclusions. The clay mineral in these oriented inclusions shows aggregate polarization in the cleavage, indicating some recrystallization associated with flattening.

The difference in behaviour between the granular clay aggregates and the mudstone inclusions is believed to be due to the contrast in grain size. The smaller, rounded particles can accommodate themselves into the pattern of deformation by intergranular movement in the matrix, but fragments larger than the quartz grains must undergo bulk strain by flattening to the same or greater extent as the bulk strain of a granular aggregate of equivalent size.

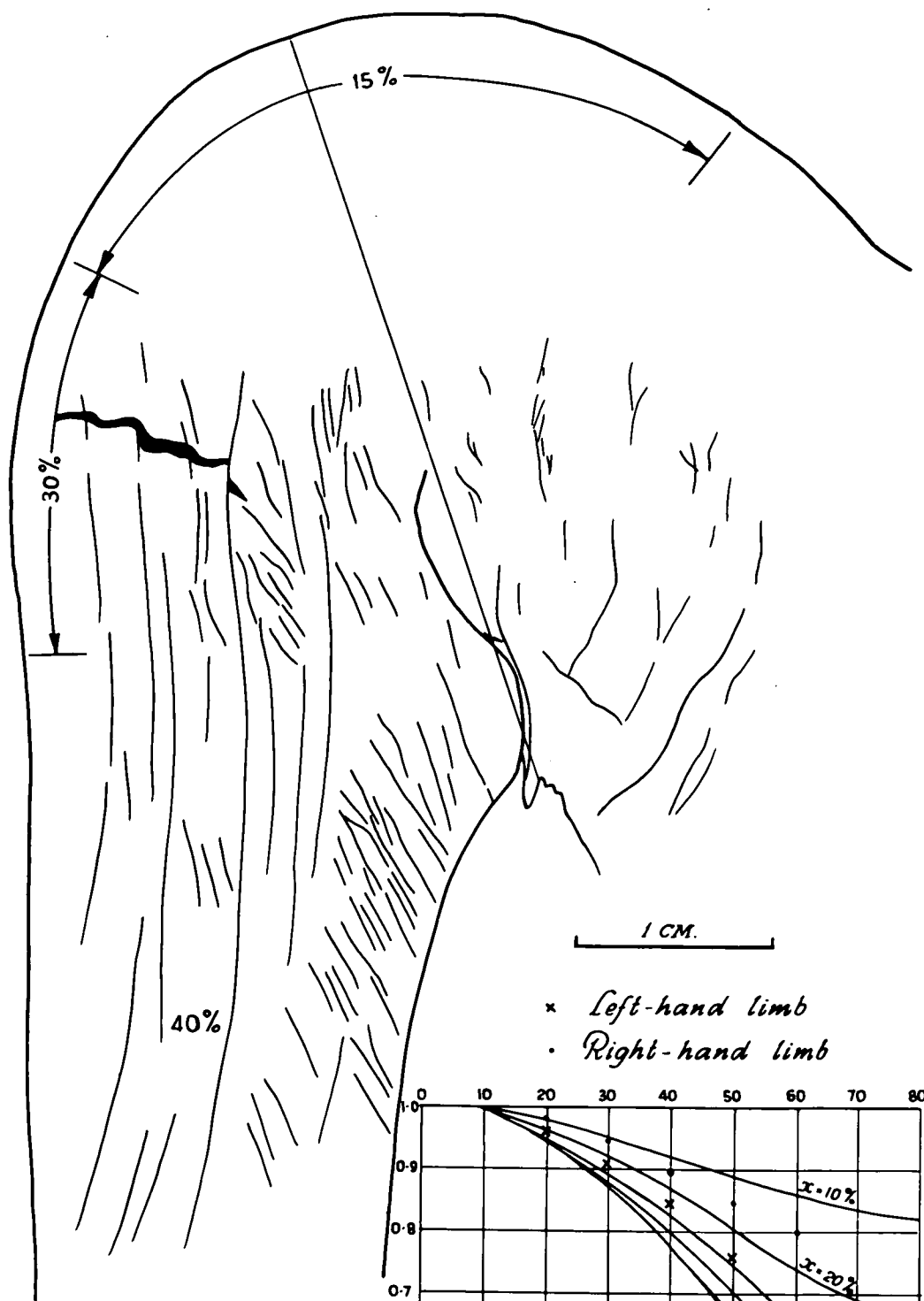
Relation between microfabric and flattening :

An example has been described suggesting that the sandstone cleavage is related to the flattening stage of fold development. An additional example, described below, shows that a relationship exists between flattening, granular deformation,

and cleavage development. The example (33317) is a tight fold in an arenite layer approximately two centimeters thick, and comes from the core region (Cooee Hinge) of the large recumbent Cooee syncline. A composite photomicrograph of a large thin section of the fold is reproduced in Plate 30, and a tracing is given in Figure 47. The outer arc of the fold, which has peeled away during the slide making, was originally marked on the glass slide and is re-drawn on the photograph.

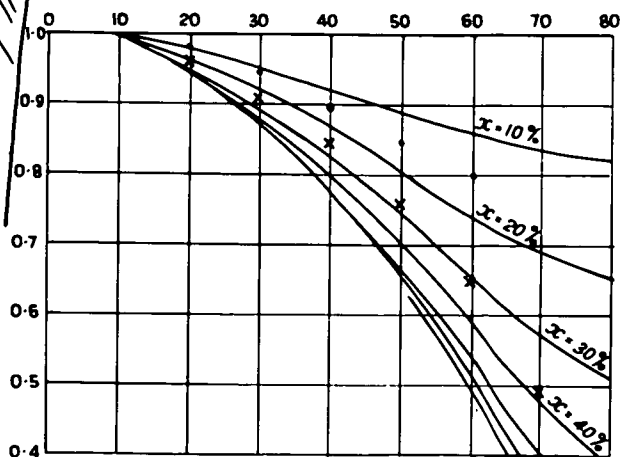
An analysis of this fold by Ramsay's method shows that flattening is not uniform. For this purpose, the trace of the axial plane is taken as the direction of the radius vector of the circumscribed circle which is tangential to the outer arc, and centred on the nick-point of the inner arc. The radius is then the apical thickness of Ramsay (1962).

The values of flattening, constructed on Figure 47 are shown. In the right-hand limb, the curve is a smooth plot corresponding to 15% flattening. However, the left-hand limb follows the 25% curve up to inclinations of 45° . At 60° it exceeds the 30% curve and at 70° it exceeds the 40% curve. According to the visual comparison method of Williams the overall picture is similar,



**DIFFERENTIAL
FLATTENING IN A
FOLD FROM OCEAN
VISTA**

FIG. 47.



but differs in detail, as indicated on Figure 47. Uniform flattening of 15% is indicated across the crest as far as the bulge on the left-hand limb. Beyond the bulge is a slight concavity, and the flattening increases up to 30%. This discrepancy may be due to one of several factors, (a) the pre-flattening fold profile may not have been a true circular arc, (b) the nick-point may migrate down the axial-plane due to appression of the limbs which would give too large a value to the appial thickness, and thus invalidate Ramsay's method, and (c) the presence of cross-bedding, which is suggested on the left-hand limb in Figure 47, and can be seen elsewhere in the specimen. This would affect the graphical method which assumes uniform thickness of the flattened layer. However, the overall picture of the flattening increasing from small values on the crest up to large values on the limb is considered to be realistic.

In thin section (Plate 30) there is a visible relation between the differential flattening and the microfabric. In the crest of the fold near the outer arc the rock appears undeformed. The bedding lamination is faintly visible as layers of varying grain size and the texture has a disrupted framework of quartz grains in an

argillaceous matrix. There is no sign of cleavage in the matrix, and no granulation or fracturing of individual grains. Detrital muscovite remains oriented in the bedding.

Toward the middle of the layer, widely spaced, irregularly branching cracks appear, radiating from the nick-point of the inner arc. These cracks are expressed as zones of brown iron oxide staining, and recrystallization of secondary mica without aggregate polarization. The areas between the cracks are without a perceptible imposed fabric.

At the nick point there is penetration of the slaty material into the silty bed, forming flame-like piercements. Although it appears from the photograph that these piercements go well into the sil^tstone, this is not so, as they are merely starting points of tight bundles of anastomosing sandstone cleavage.

Coming down the left-hand limb, where the flattening is greater, a typical sandstone cleavage is present. This is a closely spaced, penetrative cleavage in which shreds of sericite anastomose and enclose the individual quartz grains. The sericite in these shreds and in the lenticles of matrix has a preferred orientation sub-parallel

to cleavage. This cleavage is distinct from the diffuse bifurcating cracking in the crestal region.

On the left-hand limb there is also a cleavage parallel to the bedding lamination. Displacement of the inner-most laminae away from the core of the fold is shown by radial quartz veins. Microscopically, this bedding cleavage is due to thin zones of anastomosing shreds. Within the zones of bedding cleavage, the anastomosing shreds define two acutely intersecting surfaces corresponding in direction to the bedding cleavage and the oblique cleavage. The individual sericite flakes are commonly arranged en échelon with their basal planes in a mean direction which lies between the trend of bedding cleavage and oblique cleavage. The relationship between bedding cleavage and oblique sandstone cleavage suggests that both surfaces developed synchronously.

It is tempting to explain the sandstone cleavage simply as planes of high resolved shear stress, along which there has been movement.

However, there is no evidence of movement on the sandstone cleavage, even though displacement is shown on the bedding cleavage. Furthermore, assuming the plane of flattening is correctly identified, it is seen that the cleavage in the

arenite is parallel or only slightly acute to the flattening plane, and therefore cannot be planes of high resolved shear stress.

The strain pattern after the initial buckling is therefore non uniform, with the crestal portion behaving as a rigid rod, responding only slightly to the applied stress, whereas on the limb there is progressively ^{greater} flattening away from the rigid crest. If this differential flattening takes place across a sharp zone a rupture may occur and will be expressed as oblique shear joints with small displacements. If it is dispersed across a broader zone, a bulge in the outer arc will occur.

Theories on the origin of "fracture" cleavage

A diversity of mechanisms for the origin of fracture cleavage can be found in the literature. This in part reflects the different usage of the term. Some writers believe it to be entirely independent of mineral orientation and hence mechanically distinct from slaty cleavage, while others consider that it is gradational into slaty cleavage. The suggested origins of fracture cleavage fall into three categories.

Wilson (1946) and Campbell (1951) suggested that fracture cleavage is a set of mechanically generated slip planes directly related to the interbed shear during concentric folding of the

competent layer. An objection to this (de Sitter 1956, p.98), is that the cleavage is just as strongly developed, if not more so, in the crests than in the limbs. Nevertheless, Maxwell (1962) has persisted with such an origin on account of the bed-delimited nature of the fracture cleavage. To overcome the above difficulty, he postulated a sequence of cleavage formation on the planar limbs of the major fold, and the subsequent development of the minor folds by the migration of the fold form along the bed.

In another category are invoked the movements in the buckled bed itself rather than the interfbed shear. Hills (1963, p.306) suggested that some fanned fracture cleavage may form initially as tension fractures at right angles to the bed, and then be rotated along with the bedding to the present position. Similarly, Knill (1960, p.320) suggested that where the amount of fanning is small with respect to arc length, the fracture cleavage initiated as tensional cracks during concentric folding but the rotation away from the axial plane was counteracted by a later phase of similar folding (flattening). This mechanism, while adequate for a close-spaced planar jointing independent of mineral orientation, does not explain the sandstone cleavage in the Burnie Formation. The geometrical relationships that

have a bearing on the mechanism of Knill were briefly described by Gair (1949), who found that fan cleavage intersects bedding at 90° in the fold crest, but this angle decreases down the limbs when the thickness decreases.

Voll (1960, p. 551) following the German writers Hoeppeener, Mosebach and Wunderlick, maintained that some fracture cleavage forms as acutely intersecting shear fractures at an early stage in flat-lying layers. The angle between shear joints and bedding changes progressively with rotation of the fold limbs due to flexural flow parallel to bedding. Subsequent development of mica fabric may then be due to selection of suitably oriented seeds.

Recently, Williams (1965) has suggested that for some flattened concentric folds, ^{the cleavage} /in the crest initially formed in the radius of the concentric fold in the competent bed. This origin of the cleavage is based upon the property of dilatancy in which zones of loose texture pass through the layer with continued bending. Williams continues (p.234) "clearly the zones of loose packing of the sand grains of the framework will be the sites of the planes of cleavage." The phenomenon of dilatancy appears to be a process that must be considered during the deformation of granular aggregates, except perhaps those with an

abundance of matrix with a disrupted framework. However, it appears that the sandstone cleavage in the Burnie Formation is related to the flattening phase of fold development, rather than the initial concentric folding. It should be added that many purely concentric folds of the P3 generation are developed in the same rock type as the P1 folds, and have no cleavage.

The third category arises from the view that fracture cleavage and slaty cleavage are genetically related. Thus de Sitter (1956, pp.97-98, p.214) maintained that fracture cleavage marks the planes of extension resulting from a process of flattening of microlithons combined with a pure shear. The two types of cleavage differ in the degree of recrystallization, and the width of the microlithons.

Development of sandstone cleavage in the Burnie Formation

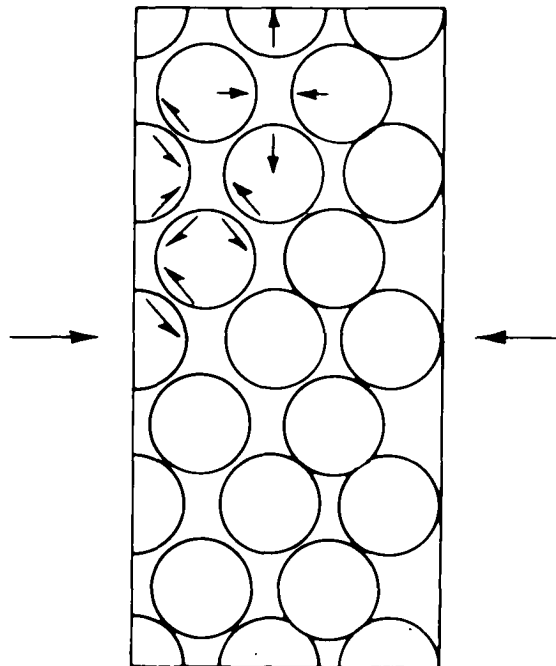
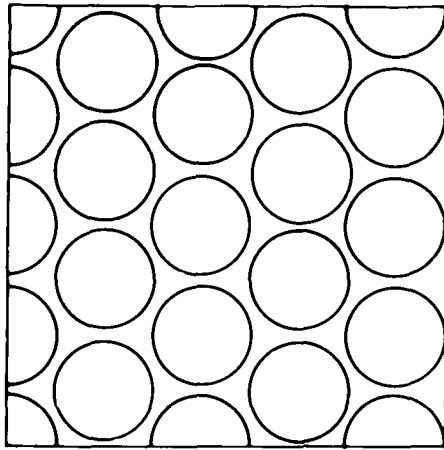
The relationships between the fold form, the cleavage, and the deformation at the granular scale allows some comment to be made on the development of sandstone cleavage. The detailed movement picture of deformation at the granular scale is heterogeneous, with the equant quartz grains behaving as rigid particles and the greater part of deformation occurring within the argillaceous

matrix. The clay aggregates, which are of similar size to the quartz grains accommodate themselves into this pattern of deformation and generally remain intact. The flakes of detrital muscovite, which are larger than the mean grain size show evidence of bending and splintering.

This type of deformation may still be regarded as statistically homogeneous in the usage of Turner and Weiss (1963, p.366), because the microscopic strain discontinuities are statistically the same in all samples. Thus the values of flattening quoted here represent the estimate of mean strain of the statistically homogeneous deformation on scales larger than the thin section.

The movement picture envisaged during flattening involves a movement together of the grains in the flattening direction and drifting apart of the grains within the flattening plane, with the grains being cushioned by the interstitial matrix. This process is illustrated diagrammatically in Figure 48, where a granular aggregate having a disrupted framework is flattened by 30%.

Intergranular shearing occurs between two grains that are approaching each other and the pinched matrix is intensely sheared. A mean plane of more intense shearing is formed in the flattening plane,



30% flattening of a granular aggregate initially with a disrupted framework.

FIG. 48

along which the individual displacements are opposite and mutually cancelling. Thus, there is no discrete plane of shearing with a consistent sense of movement, but rather statistically planar surfaces which anastomose and enclose the rigid grains. If the clastic grains are considered to be equant, a similar movement will occur in the third dimension. This would correspond to movement apart of the individual grains along the fold axis, but may be suppressed by confinement of the surrounding material. In the absence of abundant matrix there is a limit beyond which flattening can only proceed by suturing, fracturing and granulation. Up to this point the granular aggregate would be dilatant.

It appears that the intergranular shear zones of pinched matrix are the zone of initial reorientation of the clay mineral. The exact mechanism of reorientation is uncertain, but the situation may be analogous to the experiments of Weymouth and Williamson (1953) who repeated the clay deformation experiments of Riedel (1929).

In the experiments, small shear planes were produced in plastic clay by the application of a simple shear. Weymouth and Williamson (1953) observed that the shear was a discrete zone of movement, filled with clay flakes in which the basal planes were oriented subparallel to the shear zone.

They suggested that "the extreme fineness of the clay particles enhanced their ability to mechanically reorient themselves even where the movement was slight." In the rocks under discussion, the amount of movement along the intergranular shear zones is of the order of a single quartz grain radius which is large compared with the size of the individual sericite flakes of the matrix. In this way, a fabric similar to that shown in Plate 28-a may be produced.

The next step, in which the anastomosing shreds of secondary sericite are produced, may be due to mimetic crystallization along the statistical planes formed by the discontinuous zones of intergranular shear. In this respect it is noteworthy that Weymouth and Williamson (1953) observed that the individual flakes in the infillings of the shear zones were inclined at up to 15° to the shear planes. A similar feature is present in the long shreds of sericite in the arenites under discussion.

Certain petrographic features of the cleaved arenite may give some indication of the condition of the rocks at the time of deformation. Experimental deformation of sand and sandstone by Maxwell (1960) and Borg, et al, (1960), show that the compaction and deformation is purely cataclastic

involving fracturing and granulation of the quartz grains. However, in the arenite of the Burnie Formation, fracturing is at a minimum and granulation is uncommon, even in rocks flattened by up to 40%. Subsequent experiments by Handin et al (1963) show that even under high confining pressures of 2 kilobars (27,000 feet of sediment), purely intergranular deformation, without cataclastic effects, is possible by reducing the effective confining pressure. This is achieved by deformation ofunjacketed specimens, or geologically speaking, a porous water-filled aggregate.

Hubbert and Rubey (1959) have discussed the theory of such a two-phase system, and show that this behaviour is explicable by the Coulomb theory of failure,

$$t = t_0 + (S_n - P_n) \tan \phi$$

where t is the shear strength of the rock, t_0 is a constant corresponding to the cohesion of the granular aggregate, $(S_n - P_n)$ is the normal stress less the internal pore pressures, and ϕ is the angle of internal friction. As the internal pore pressure increases, the shear strength rapidly diminishes to a limit where it is equal to the cohesion of the rock. Stress concentrations are then sufficient

to cause rupture, and intergranular movement may then occur after the initial breaking of intergranular bonds provided by the matrix.

Additional petrographic features also suggest that the rock possessed mobility during deformation. These features include rotation of the large detrital muscovite flakes through a granular material, and the apparently rigid behaviour of small granular clay aggregates compared with the plastic behaviour of larger clay fragments. The presence of brown iron staining of the sericite in some of the more conspicuous cleavage shreds may also indicate an abundance of interstitial water. Hills (1963, p.299) also notes features which may indicate the importance of interstitial water, as also does Maxwell (1962) who applied the concept of high internal pore pressure to a theory of slaty cleavage. Similarly, Turner and Weiss (1963, p.459) state that the viscous behaviour of recrystallizing slate may be achieved by pore pressures which equal or exceed the load pressure.

When applied to granular beds containing abundant matrix, this process may explain the change-over from initially competent behaviour during early folding to later incompetent behaviour during the

stage of flattening and cleavage formation.

The presence of abnormally high interstitial pore pressures does not demand any special environmental conditions and does not necessarily imply that deformation occurred in unconsolidated sediment at a shallow depth of burial. Thus Rubey and Hubbert (1959) suggest that it is a natural consequence of rapid deposition of thick sections of alternating medium- and fine-grained sediments. The un-jacketed tests of Handin et al (1963) indicate that the ductility and shear strength of any rock are considerably reduced providing the rock is porous to some extent. The test specimens of sandstone and siltstone were taken from depths between 5,000 and 12,500 feet.

Anomalous pore pressures could be generated in a thick sequence of arenite and mudstone, because the loss of permeability with compaction for the finer grained beds is greater than for the coarser grained beds. Thus a pressure gradient is set up across the boundary between the slowly compacting arenite and the impermeable lutite. This effect may be accentuated by the application of tectonic pressures, which would accelerate the rate of loss of permeability of the sealing beds.

Slaty Cleavage

The slaty cleavage is generally parallel to the axial plane, but may show a slight upward fanning or an upward convergence symmetrical about the axial plane. In the folds displaying the mesoscopic asymmetry in the Blythe Overturned Belt, the slaty cleavage in the fold cores is parallel to the sandstone cleavage at the crest, and is thus slightly acute to the plane joining the inflection points. However, the traces of cleavage on the bedding are parallel to the fold axes.

Some rare examples of slaty cleavage oblique to the fold axis are found at Sulphur Creek at a point 400 yards west of Sulphur Creek headland in a dominantly slaty horizon. In this example, the dominant slaty cleavage is oblique by 15° to the axial plane, and intersects with the axial plane in a line which is perpendicular to the fold axis. Therefore no anomalous cleavage is seen in the fold profile, but lineation oblique to the fold axis is seen on bedding planes (Plate 26-b). An earlier subordinate cleavage is parallel to the axial plane to this fold. The fold axis and the earlier axial-plane cleavage exhibit a slight curvature about a vertical axis, whereas the later cleavage is planar.

This configuration, although unusual, is compatible with the concept that cleavage developed continuously during the formation of the minor structures. It is suggested that the folds with oblique lineation were rotated out of the position of formation, perhaps by differential flattening on a broad scale, and a continuation of the development of slaty cleavage in its initial attitude resulted in the oblique slaty cleavage.

In thin section (specimen 33318), the slates have the fabric of a typical slate with a strong planar orientation of basal planes of mica. No precise determinations have been made on the mica but it is optically similar to the sericite found in the cleaved arenite. Individual mica flakes are clearer and slightly larger (0.005 - 0.02 mm) than the sericite present in the uncleaved mudstone. In contrast with the cleavage in the arenite, there are no discrete planes of cleavage, but merely a statistical plane of parting due to the alignment of fine micas. No microscopic segregation parallel to the cleavage is present in the majority of the slate, however an incipient segregation does occur in the Doctors Rocks area, where the P1 cleavage

is believed to grade into the foliation in the Keith Metamorphics.

In the purer slates, the accessory constituents include detrital quartz (0.02mm), detrital muscovite, tourmaline, rutile and zircon, which are dimensionally oriented in the cleavage. The possible presence of a linear dimensional orientation perpendicular to the fold axis, so often reported in slate has not been investigated, but a lineation perpendicular to the bedding-cleavage intersection is commonly seen. Minute cross-fibre aggregates of muscovite and chlorite occur.

With increase in proportion of detrital quartz (as the mudstone passes into siltstone), the cleavage develops into a finely anastomosing sandstone cleavage, with the detrital quartz grains assuming a micro-porphyroclastic habit.

Theories on the origin of slaty cleavage

Axial-plane slaty cleavage is generally considered to form perpendicular to the principal compressive stress. This view is due mainly to the early British writers, who proposed various mechanisms. Thus, Sharpe (1846) believed the whole rock mass, including the constituent particles, underwent a compression with associated extension in the plane of the cleavage. Later, Sharpe (1849), suggested

that much of the preferred orientation was due to mimetic growth of mica. Sorby (1856) advocated mechanical rotation of rigid mica flakes in a plastic matrix into a plane normal to the compression. Harker (1885) followed Sorby, but later (1932) emphasized the role of syntectonic crystal growth especially of mica, and outlined the concept of stress minerals. These workers, and others, recognised that fossils (Haughton, 1856) and reduction spots (Sorby, 1908) were flattened in the cleavage plane.

Much evidence has been presented showing that slaty cleavage, as a broad approximation, forms normal to the direction of shortening. The mere presence of axial-plane cleavage in folds having the characters of both concentric and similar folds is strong evidence, (Swanson, 1941; Morris and Fearnside, 1926). Furthermore, the theories of elastic and elastico-viscous buckling of Currie, Patnode and Trump (1962), Ramberg (1962, 1963) and Biot (1961) demand that the axial plane of an elastically induced buckle is perpendicular to the direction of compression.

Pre-cleavage strain indicators have been used to show that the cleavage is a plane of flattening. The use of fossils dates back to Haughton (1856) and more recently Breddin (1957)

has demonstrated triaxial strain with extension in both directions in the flattening plane. Cloos (1947) showed that oolites were flattened in the cleavage plane and that cleavage appeared with 30% flattening. Similarly in deformed conglomerate, the greatest and mean diameters are aligned in the foliation, for example, Oftedahl (1948) and Burns (1964).

The correlation of the plane of flattening with the plane perpendicular to the principal maximum stress is supported by the symmetry arguments of Turner and Weiss (1963, p.454), the experimental deformation of Yule marble (Griggs, et al, 1960) and from the syntectonic recrystallization of quartzite (Carter, et al, 1964).

Consequently, the view that the plane of slaty cleavage is the plane of flattening, normal to the direction of lateral shortening has become generally accepted in recent years, for example, Gougel (1962, p.42), de Sitter (1956, P.97, p.215), Hills (1963, p.287) and Turner and Weiss (1963, p.447-460).

Most discussions are now centred on the orienting mechanisms responsible for the preferred orientation in slate. The early ideas of mechanical rotation, syntectonic crystallization or mimetic recrystallization have been mentioned. More than

100 years of discussion and speculation has not contributed greatly to a full understanding of the problem. Of the more notable recent writers advocating a dominantly mechanical origin for the orientation of mica, Maxwell (1962, p.300) advanced a mechanism involving orientation by flowage parallel to cleavage during collapse of an open network of clay plates and fibres with water filled interstices. During deformation, the load is carried by the connate water allowing flattening, rotation and extensive flowage parallel to the cleavage, resulting in a near-parallel orientation of clay particles. This theory was mainly based on the argument that the mica of slate was illite and not sericite as is normally presumed, so that the slate fabric was not produced by crystallization associated with the onset of progressive regional metamorphism.

Flinn (1962, p.399) theoretically examined the development of preferred orientation of randomly oriented planes during finite homogeneous strain for different strain systems. As the deformation ellipsoid changes from prolate to oblate, the expected preferred orientation changes from a uniform girdle of (mica) poles in the plane normal to the maximum axis, to a point concentration of poles at the minimum axis. Although such orientations are to be expected, the degree of preferred orientation,

even at high strains, does not approach the high preferred orientations found in most slate. Thus Flinn (1962, p.424) concluded that mechanical rotation during finite homogeneous strain does not constitute a suitable orientating mechanism.

Most writers follow Leith (1913) in attributing a dominant role to crystallization. Whether this is syn-tectonic or post-tectonic is one of the problems. Hills (1963, p.304) and de Sitter (1956, p.215) imply growth of micas in some mechanically formed slip plane. Another point of discussion is the role of mechanical rotation in the crystallization process. Behre (1933, p.45-56) considered the selective growth of sericite from initially favourably oriented micas at the expense of others not so well disposed to be important. Williams (1961-b) suggested rotation of the initial clay minerals into planes of laminar flow within the incompetent beds, followed by recrystallization in these planes.

Collette (1956) outlined a theory of slaty cleavage based on the thermodynamics of the growth of mica in a stress field. This is an elaboration of Riecke's Principle, accounting for lattice orientation as well as dimensional orientation. The fine mica responds to infinitesimal elastic stress by solution in some crystallographic directions and growth in other directions, depending on the relationship

between the stress field and the lattice orientation.

Some of the questions of slaty cleavage formation have been outlined above, and it is doubtful whether there is any petrographic evidence that may provide answers. It appears that slaty cleavage in the Burnie Formation formed normal to the direction of flattening, and that sandstone cleavage and slaty cleavage have a similar origin. It was considered that mechanical rotation was the important initial process in the formation of sandstone cleavage whereby the mineral orientations were produced by a heterogeneous strain pattern controlled by the granular material. It is possible that the high degree of preferred orientation in the slate may be achieved by a heterogeneous strain pattern on an even smaller microscopic scale, in which the individual clay-mica flakes behaved independently in a mobile flowing medium. In detail this may be achieved by the process outlined by Maxwell (1962).

Relation of the P1 cleavage to the foliation in the Keith Metamorphics

It is suggested that the P1 cleavage in the Burnie Formation is equivalent to the foliation in the Keith Metamorphics. The contact between these two assemblages is covered by Permian sediments and Tertiary basalt. However, it appears from the inland exposures that the rocks just to the east

of Doctors Rocks lie within one or two miles of the contact. Various evidence suggests that this boundary is gradational like the western boundary of the Keith Metamorphics. Thus, the P₄ cleavage which is strongly developed near Doctors Rocks diminishes to the east and disappears at Ocean Vista. Similarly, the P₁ cleavage is strongly developed at Doctors Rocks and five miles to the east at Coeee it vanishes.

The transitional nature of the P₁ cleavage has not been examined in detail. Present information suggests that the grade of metamorphism increases westward toward the Keith Metamorphics. At Ocean Vista the sandstone and slaty cleavage has the normal mesoscopic and microscopic aspect. Further to the west at Somerset the cleavage becomes dominant over the bedding, until near Doctors Rocks it is possible only to recognise remnants of bedding in the cores of the P₁ folds. The cleavage then loses its bed-delimited character and takes on the form of a planar penetrative foliation with a distinct but discontinuous banding parallel to the foliation.

In thin section, (33309, Plate 29-b) these rocks are composed mainly of fine-grained micaceous material with disrupted layers or pod-shaped lenses of quartz siltstone. The lamination appears to be

a transposed surface with bedding at a small angle to cleavage. The non-crystalline nature of the quartz siltstone pods indicates that this banding is not a metamorphic segregation. The micaceous ribbons show strong aggregate polarization and contain minute isolated quartz grains that are flattened in the foliation and wrapped by the mica flakes. Elongate micro-augen of cross-fibre chlorite (penninite) are wrapped by the mica foliation. The sericite in the matrix of the siltstone pods also have a preferred orientation. The quartz grains generally are small (0.02mm) and appear to be the original grains showing few effects of flattening or granulation. In a few of the smaller pods of siltstone, the grains are larger (0.2mm) and are interlocked. These may represent local patches of ^{re}crystallization.

Concentric cleavage

A cleavage has been observed parallel to bedding in the core of the large mesoscopic anticline between the two main layers of pillow lava at Sulphur Creek, (see Figure 57). It has been observed only in this horizon which is the thickest competent horizon in that area. The cleavage is found near the outer arc of the fold, and the usual axial-plane sandstone cleavage is found nearer to the

core of the anticline. Mesoscopically, the concentric cleavage is indistinguishable from the axial-plane sandstone cleavage, being a penetrative, closely spaced, finely anastomosing plane of difficult parting. It is readily distinguished by its relation to the fold because it lies parallel to bedding or inclined at very small angles to bedding.

De Sitter (1964, p.292) mentions a similar feature which he calls concentric cleavage. It has the property of being restricted to the argillaceous beds, where it is at low angle to bedding and even curves into the bedding plane when approaching a competent bed. De Sitter also notes that it is due to a parallel arrangement of mica.

The concentric cleavage at Sulphur Creek is different from the above example, because it is found in the competent arenite bed and is not necessarily related to the parallel arrangement of micas. Microscopically, (Plate 31-a), it is different from the axial-plane sandstone cleavage because the cleavage planes are shear planes. The cleavage is due to fractures which although they are anastomosing, are not completely controlled by the larger detrital quartz grains. These planes chop through the quartz grains, producing fracturing and border granulation. Detrital muscovite is not

preferentially oriented into the cleavage.

Both concentric cleavage and axial-plane cleavage can be found in the one thin section, (Plate 31-b). The axial-plane cleavage shreds have an en echelon nature, splaying out into the concentric cleavage plane. This suggests a contemporaneous development of the two cleavages.

The presence of concentric cleavage in a large well-rounded anticline in a thick competent horizon involving a small amount of flattening (10% according to the visual comparison method), points to an origin related to bedding-plane shear accompanying concentric folding. During concentric folding, bedding-plane shear may be dissipated along discrete shear planes which follow inherent lithological planes, or by more penetrative movement on a scale extending down to the individual quartz grains. These two types of deformation have been termed flexural slip and flexural flow by Donath and Parker (1964). The reason why this concentric shearing is reflected by cleavage and not the usual flaggy parting may be due to the thick massive bed having no inherent lithological layering that may act as potential planes of shear.

ASSOCIATED STRUCTURES

Concentric Shears

Concentric shear joints are well developed in the competent arenite layers. At Somerset, where flattening is greater, the bedding fissility is obscured by the strong development of axial-plane cleavage. In the Blythe Overturned belt the concentric shears are expressed by a closely spaced flaggy bedding fissility such as on the left-hand side of Plate 32-a. The fissility is most strongly developed in the straight zones on the outer limbs of the coupled folds, but is also present on the common limb.

Rare examples of slickensides on the concentric shear joints perpendicular to the fold axis have been observed, at a point half a mile west of the mouth of the Blyth River. Slip along these planes can be demonstrated at several localities (for example, Sulphur Creek headland), by offsetting of quartz veins which penetrate along rotational joints. The sense of displacement is that normally presumed for concentric shear joints, that is, the beds nearest the centre of curvature moving out of the fold cores.

Although much of this fissility occurs in weakly laminated or massive arenite, and therefore is

mainly mechanically generated, a control by inherent internal sedimentary lamination is apparent.

Commonly, the concentric shears are expressed as undulatory planes of parting which follow the bulbous basal surfaces of the festoon cross-bedding.

Plate 32-a is an example of festoons strongly modified by concentric shears, in which the concentric shears bifurcate to enclose lenticular bodies.

Referring to the same examples, Burns (1964, p.148) calls these "pseudo-boudins", and noted that the axes of the swells pitch in two directions approximately at right angles in a "chocolate-tablet" structure.

This arrangement is to be expected considering the initial form of the festoon scoops.

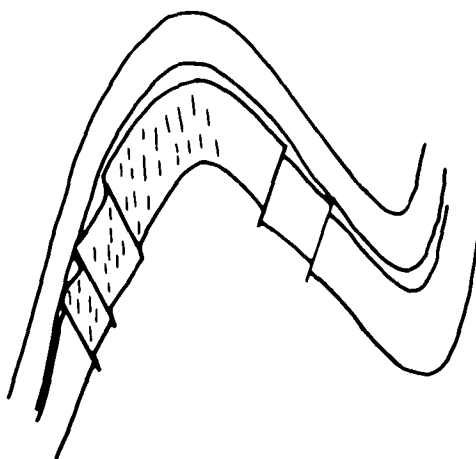
Oblique shear joints

On the limbs of ⁿ many of the folds at Sulphur Creek are oblique shear joints which cut and displace such structures as bedding, concentric shears and cleavage, (Plate 32-b). These joints are generally inclined at from 45° to 70° to the bedding, and are seldom at right angles to the bedding. It is therefore more appropriate to term them oblique shear joints in the meaning of de Sitter (1956, p.100), rather than rotational joints, as by Burns (1964, p.148).

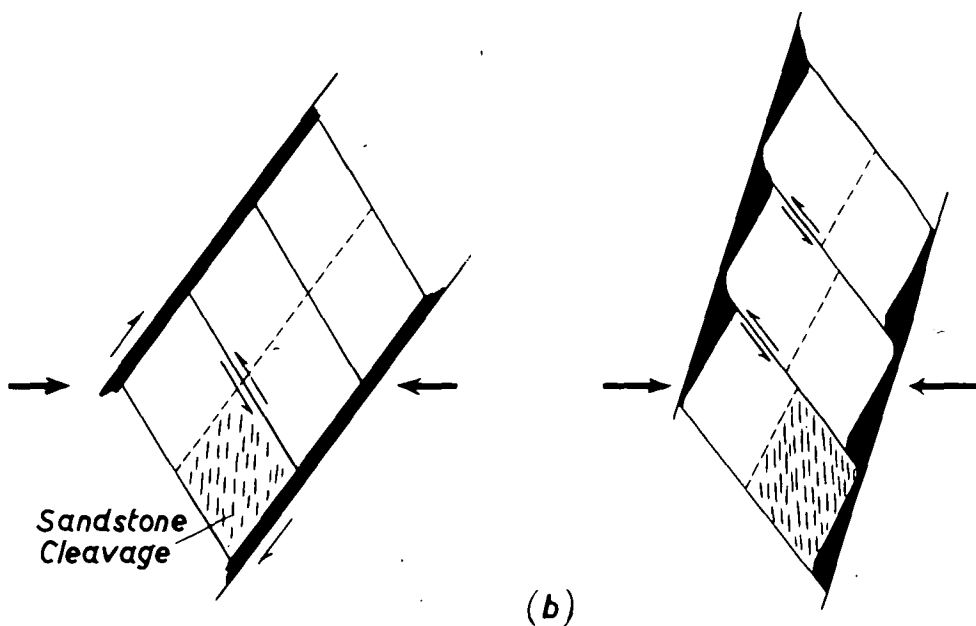
Oblique shear joints are spaced about 9 inches apart and cut across beds up to 30 inches thick. In thicker layers they tail out along concentric shears, (Plate 32-b). In the incompetent material they initially cut the slaty cleavage, but rapidly turn into the cleavage. Displacement along the joints is often accompanied by rotation, leading to incipient boudinage.

The oblique shear joints maintain a fixed relationship to the fold, (Figure 49-a). They occur on both the outer and the common limbs of the fold couplet, where they have opposite displacements, and in profile, fan more or less symmetrically about the axial plane. The joints are inclined at 30° - 40° to the axial plane and intersect with the bedding in a line parallel to the fold axis. The sandstone cleavage bisects the acute angle of the rhomb formed by the concentric shear joint and the oblique shear joint.

The features of the oblique shear joints indicate that they are the expression of late-stage shortening under the same stress field operating during the development of the earlier P1 structures. The shortening was achieved, not by flattening in the usual sense, but by rotation of the joint blocks (Figure 49-b). The joint plane can be



(a)



(b)

a. Generalized disposition of oblique shear joints

b. Development of oblique shear joints and boudinage

interpreted as a surface of high shear stress with the concentric shear acting as its conjugate pair. The principal stress direction is the bisector of the obtuse angle of the rhomb which is perpendicular to the sandstone cleavage and approximately perpendicular to the axial plane of the nearby fold. It should be noted that in Figure 49-b, there is an apparent clockwise rotation of the joint rhombs, but actually there is anticlockwise rotation, although this is less than the anticlockwise rotation of the ⁱⁿconfining boundaries. Thus, the formation of oblique shear joints allows further tightening of the fold during the late stage of folding.

Boudinage

Boudinage occurs at Sulphur Creek in some of the arenite layers which are from 6-12 inches thick and encased in thicker layers of mudstone. Plate 33-b illustrates a typical example, as also does Plate 16 of Burnet (1964). The boudins have a length-to-width ratio varying from 7 : 1 down to 4 : 1, which is generally less than that found by Ramberg (1955). In many cases it is found that cross-bedding controls the position of the nodes. Vein quartz is present in the nodes. As a rule, the boudin axis is parallel to the fold axis, although deviations do occur on account of the "chocolate-

tablet" arrangement of the cross-bedding.

There is no clear evidence relating to the stage of development of boudinage. In some cases it is clearly related to the stretching of the limbs during the formation of the disharmonic carinate structures, (Figure 46) and is therefore a late structure.

Related Faults

A system of faults following the P1 cleavage is present between Round Hill Point and Sulphur Creek headland (Figure 57). These consistently trend 223° and displace east-side-north. These faults break cleanly through the arenite layers but commonly in the slate they are expressed as a closely spaced planar sheeting which may be termed a type of fracture cleavage.

Because of the slightly convergent nature of the slaty cleavage, the faults intersect very acutely to the cleavage producing "pencil slates."

There is a multitude of smaller faults, unable to be placed on the map scale of Figure 57, which are related to the main system of faults mentioned above. These are axial-plane break

thrusts in the crests and oblique faults on the limbs. The oblique faults may have been initiated by the oblique shear joints. The sense of displacement is not consistent; for example, Plate 34-a shows an anticlinal core that has been displaced upward with respect to the remainder of the fold. These smaller faults are late-stage structures which cut across the cleavage.

Another type of fault which is commonly found between Blythe Heads and Howth is one resambling a low angle thrust; however the steeply inclined attitude and the oblique slip prevent it from being a thrust in the nomenclature of Anderson (1951), and is more akin to a wrench fault. These structures are shown in Figure 57, and may be likened to small disharmonic gravitational gliding folds, with anticlines cascading over each other, and the "lower-most" ⁿsyncline sheared out along the fault. The amount of displacement on this "basal" slide is probably small because, in many cases, the planar bed immediately adjacent to the "basal" slide bends around and encloses the sheared out syncline and participates in the folding.

On the extremity of Round Hill Point is a group of faults having no close spatial relationship to any Pl mesoscopic folds. They trend 043°,

dip 70° - 80° northwest, cutting the bedding at very acute angles. Displacements vary up to 15 feet and one arenite layer is cut by two faults, tripling the stratigraphy (Plate 34-b). The faults are extremely difficult to detect in the shale horizons where they run along the bedding lamination. At this locality, the P1 slaty cleavage is absent. Accurate measurements of bedding, sandstone cleavage and fault planes have shown that the faults intersect with the bedding in a line parallel to the local tectonic lineation.

All these faults appear to be related to the P1 folds, and the direction of movement is assumed to be perpendicular to the local fold axis. In the Sulphur Creek area the P1 folds are reclined or inclined, and so the faults are oblique-slip faults.

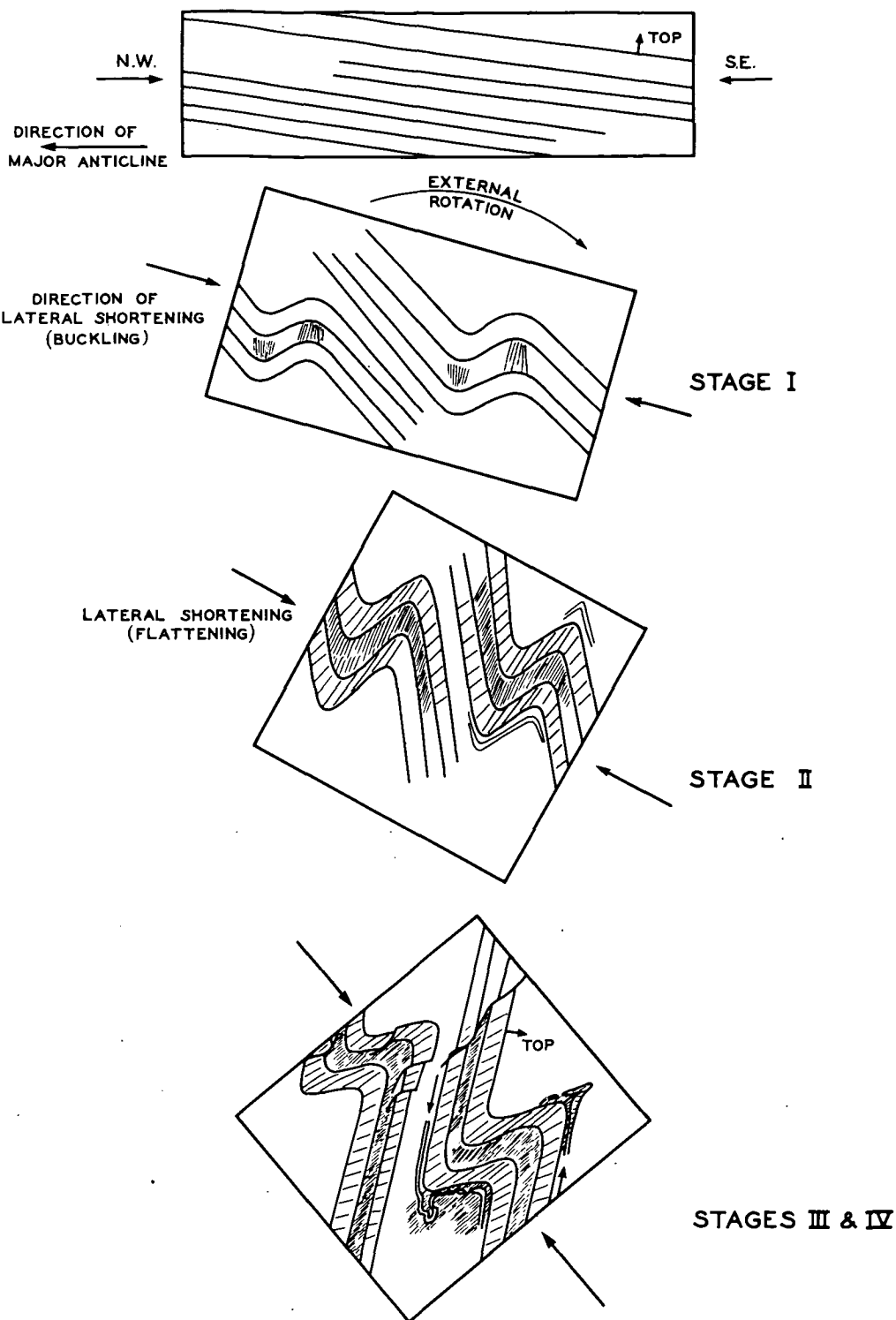
SEQUENCE OF FORMATION OF THE MINOR STRUCTURES

The mutual relations of the minor structures discussed in this chapter show that the mesoscopic structures evolved by progressive deformation in which various stages can be recognised. All these structures are grouped within the one phase of deformation because they fit into a coherent movement picture of continuous deformation about the one tectonic axis, so that on the mesoscopic scale

the symmetry remains monoclinic. This contrasts with the structures formed by the superposition of the late-generation fold structures.

The initial stage (Figure 50) resulted from buckling of the sedimentary layering, during which the competent layers deformed concentrically by flexural slip and the incompetent layers passively accommodated themselves into the interstices. The bedding fissility in the arenite layers formed at this stage, and it is probable that a localized slaty cleavage also started in the thickened cores of mudstone.

Later, flattening became important, producing the sandstone cleavage in the initially competent arenite beds and a more widespread slaty cleavage. The slight fanning nature of the sandstone cleavage may be due to a continued tightening of the folds by rotation of the limbs proceeding concurrently, with differential flattening in the core region. That much of the cleavage developed at this stage is shown by the existence of the fan and "half-fan" cleavage, the refraction of cleavage, the sigmoidal sandstone cleavage, the distortion of cleavage by structures of the later phase, (warping of sandstone cleavage by disharmonic carinate structure in Figure 47-c), and the transection of slaty cleavage



**SEQUENCE OF DEVELOPMENT OF MINOR
STRUCTURES IN BURNIE FORMATION AT
SULPHUR CREEK**

by oblique shear joints.

The third stage in the evolution is characterized by the formation of disharmonic "slip-offs" in the thinner arenite beds, and the rotation of the carinate structures out of the plane of flattening. The oblique shear joints probably formed at this stage. These structures are the result of an inhomogeneous response to lateral shortening in which extension in the profile plane occurred by various combinations of pure flattening and exaggerated bedding plane slip. Referring to Figure 50, it is seen that the bedding-planes on the straight zones of the couplet system are more favourably oriented for bedding-plane slip than are the common limbs. The net result is boudinage of the common limb, and a rotation of the carinate structures out of the plane of flattening in a direction towards the common limb. The continued development of slaty cleavage causes a localized cross-cutting relationship between the rotated carinate structures and the cleavage.

It is believed that all these structures at Sulphur Creek are the result of a unified strain history involving initial buckling and pure flattening which later became differential flattening. This was accompanied by an external

rotation associated with the major anticline to the west. It is not necessary to invoke an overall simple-shear couple acting on the sedimentary pile. The prevalence of only one plane of cleavage indicates that the direction of flattening remained constant with respect to the structures developing, although it must have rotated along with the external rotation.

In Figure 50, the profile plane underwent an additional rotation. Initially it was vertical but rotated so that the fold axis attained a significant plunge to the south west. In the final position, the profile plane strikes 328° and dips 40° northeast, and the plane of flattening strikes 025° and dips 65° northwest. The direction of flattening which appears to have been the principal maximum stress direction, is directed along a line plunging 25° toward 115° .

REGIONAL VARIATION IN STYLE

Although there is a general uniformity of tectonic style of the P1 structures over the whole of the outcrop belt of the Burnie Formation, there are regional variations. A regional variation due to spatial segregation of the mesoscopic folds has been described (Chapter 10), whereby the minor folds are confined to the major hinges and overturned

belts, and are absent from the intervening flat belt.

In the Somerset Overtuned Belt, the folds have a general approach to rectilinear axes (except where dispersed by later folding), and show uniform flattening of fairly high values. There is no evidence of either the mesoscopic asymmetry or the strong disharmony, and the cleavage approaches closely to a planar axial-plane cleavage. The cleavage in this area also shows a gradational increase in fabric development toward the Keith Metamorphics. In addition, there are few faults related to the P1 fold movements. This style is probably a reflection of the more intense shearing and higher prevailing temperature associated with the Keith Metamorphics.

In the Blythe Overtuned Belt, there is a marked variation in style between the eastern and western part. In the western part the folds are regularly rectilinear with axial-plane faults, and have a slight tendency toward the mesoscopic asymmetry. Further to the east, between Howth and Sulphur Creek headland, the P1 structures are more varied in orientation, and show evidence of a protracted sequence of development involving an early buckling stage, an intermediate plastic stage and a later brittle stage.

These variations in style may be due to the marked lithological difference between the well-bedded arenite and mudstone sequence in the western part of the Blythe Overturned Belt, and the more varied stratification of the eastern part. Here the rock types consist of thickly bedded arenite horizons, varying in thickness from two feet up to 20 feet, interbedded with thicker horizons of shale up to 200 feet thick. The pillow lava has also acted as horizons of structural weakness, showing rapid changes in thickness and marked dislocation along the contacts with the sediments.

Another factor which may have had some influence is the relatively unconsolidated nature of the sediments at Sulphur Creek at the time of deformation. These rocks are at a higher stratigraphic level than those further to the west, and were thus deformed at a more shallow depth of burial. This factor is difficult to evaluate because the stratigraphic section is incomplete, being overlain by Ordovician conglomerate with marked angular unconformity at Sulphur Creek. It is not known what thickness of sediment overlay these sediments at the time of deformation, and there is a sufficient time break in the stratigraphic record for the removal of a considerable thickness. However the structural

style is indicative of an early stage of competent buckling, then the main stage of plastic deformation when intergranular fluids may have played an important part, followed by a stage of a more brittle deformation with a return to competent behaviour. This suggests deformation of relatively unconsolidated sediments following by de-watering facilitated by cleavage development.



Plate 25a P1 fold showing thinning on common limb (right-hand limb), and half-fan cleavage, Sulphur Creek.



Plate 25b P1 fold showing thinning on common limb (right-hand limb), Sulphur Creek.



Plate 26a P1 fold showing some features of the mesoscopic asymmetry, tidal island opposite Howth Railway Station.



Plate 26b P1 anticline in mudstone showing cleavage oblique to fold axis, 400 yards west of headland at Sulphur Creek.

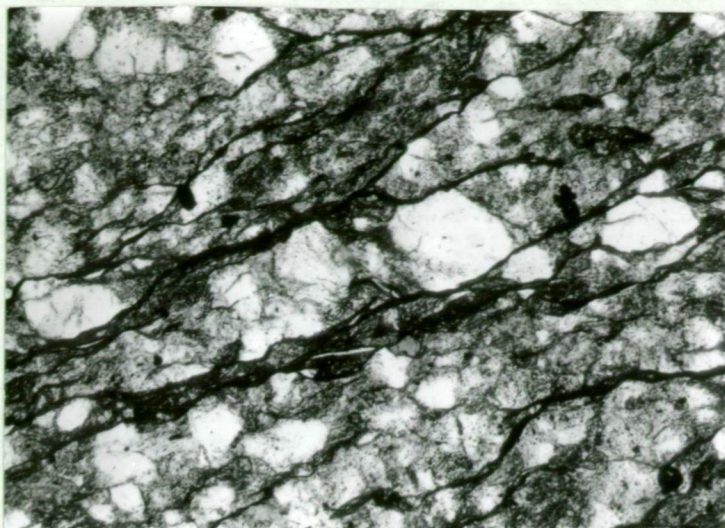


Plate 27a Photomicrograph of sandstone cleavage. Section cut parallel to fold axis and perpendicular to cleavage. Sp. 34018, x 68.

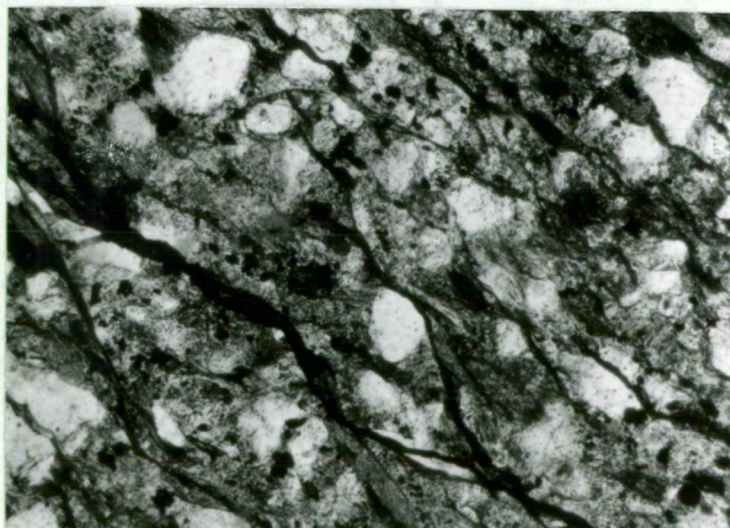


Plate 27b Photomicrograph of sandstone cleavage. Section cut perpendicular to fold axis. Sp. 34018, x 68.

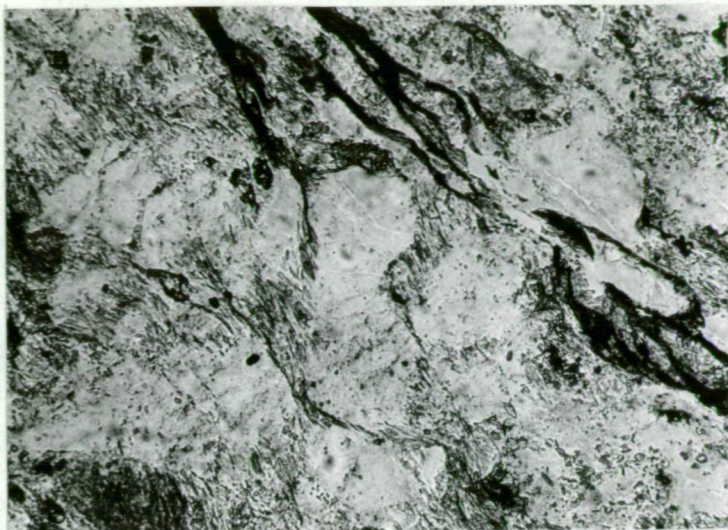


Plate 28a Photomicrograph showing detail of mica arrangement in anastomosing shreds of sandstone cleavage.
Sp. 34018, x 156.

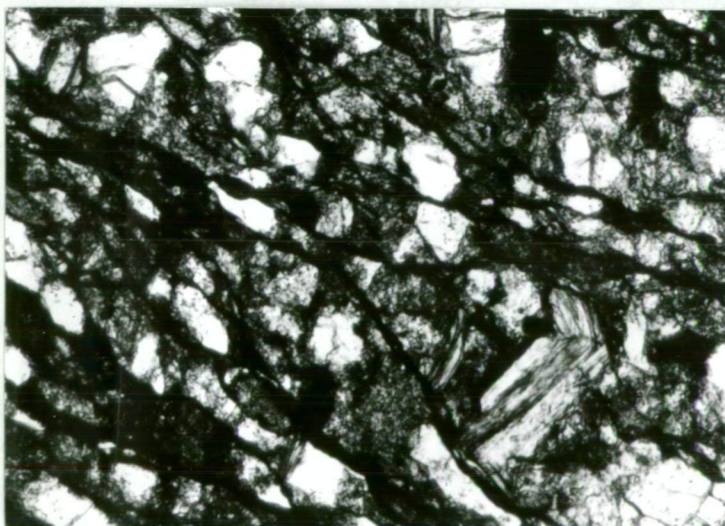


Plate 28b Photomicrograph of cross-fibre muscovite wrapped by anastomosing cleavage and showing kinking.
Sp. 34019, x 75.

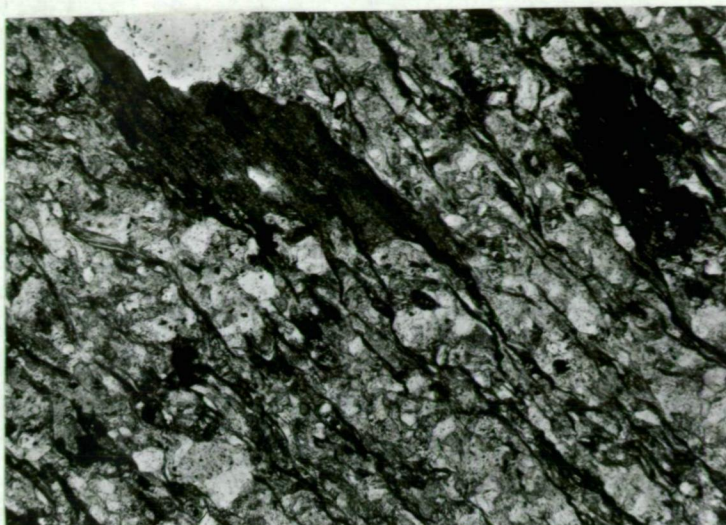


Plate 29a Photomicrograph of sheared mudstone inclusions in cleaved arenite. Sp. 33319, x 62.

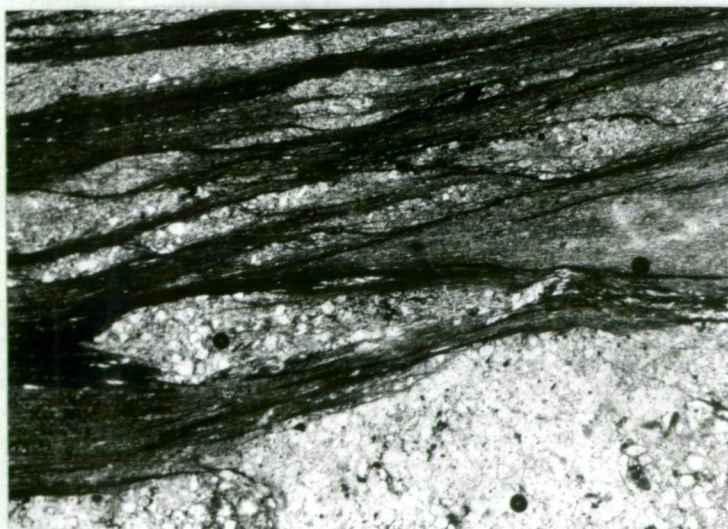


Plate 29b Photomicrograph of phyllitic siltstone showing segregated nature of cleavage, close to Keith Metamorphics, Doctors Rocks. Sp. 33309, x 22.



Plate 30 Composite photomicrograph of P1 fold showing greater flattening and cleavage development in limb than in core, Ocean Vista. Sp. 33317, x 3.

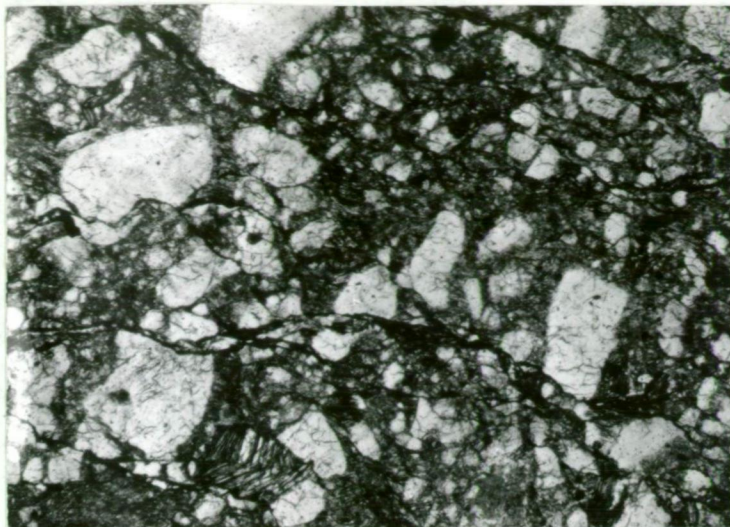


Plate 31a Photomicrograph of concentric cleavage in arenite, showing granulation and displacement of quartz grains, Sulphur Creek. Sp. 33337, x 30.

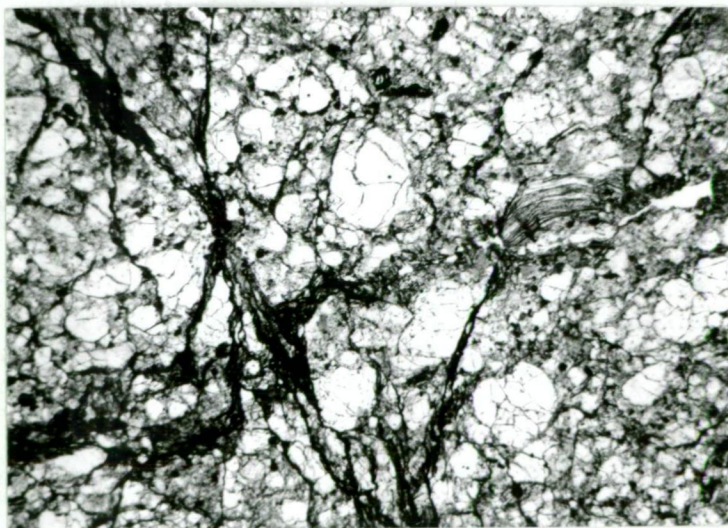


Plate 31b Photomicrograph showing interaction of concentric cleavage (sub-horizontal) and axial plane sandstone cleavage (near-vertical), Sulphur Creek. Sp. 33340, x 21.



Plate 32a Undulating concentric shears following festoon cross-bedding (top to left), headland at Sulphur Creek.



Plate 32b Displacement along oblique shear joints, top to left and next anticline is to right, headland at Sulphur Creek.



Plate 33a Oblique shear joints on outer
(left-hand) limb of P1 fold,
300 yards west of headland at
Sulphur Creek.



Plate 33b Boudinage in arenite layers,
probably controlled by cross-
bedded lenses, Sulphur Creek.



Plate 34a P1 fold with break-thrusts parallel
to axial plane, Sulphur Creek.

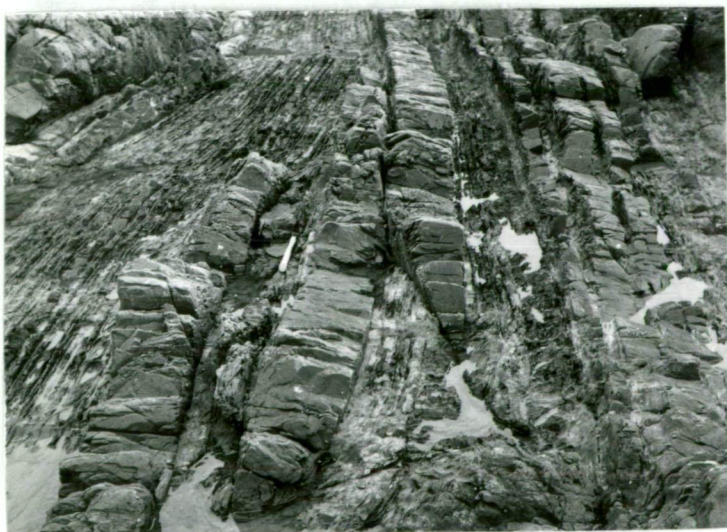


Plate 34b Triplication of section by faults
sub-parallel to bedding, Round
Hill Point.

CHAPTER 12

TECTONIC SYNTHESISMajor structural elements of northwest Tasmania

This thesis is concerned with a detailed cross-sectional analysis of the Rocky Cape Geanticline, one of the important tectonic features of Tasmania. The study shows that the geanticline is built from three major structural elements, namely, the western block consisting of the Rocky Cape Group, the Arthur Lineament, and the eastern block consisting of the Burnie Formation. Taking the northwest coast as a whole, five more major structural elements can be added (Figure 51). To the east of the eastern block is the Dial Trough, which is a northerly off-shoot of the Dundas Trough. The Dial Trough was examined in detail by Burns (1964) and considered to be an active graben filled with sedimentary and igneous rocks of a volcanic association. On the eastern side of the Dial Trough is a narrow segment of the Proterozoic basement. Still further to the east is a block of Precambrian metamorphic rocks termed the Forth Nucleus (Burns 1964, p.19). Burns suggested that this is a segmented portion of the Tyennan Nucleus of Carey (1953), which is the oldest palaeotectonic unit of Tasmania.

MAJOR STRUCTURAL ELEMENTS NORTH-WEST TASMANIA

SHOWING THE IMPORTANT
REGIONAL CORRELATIONS

0 16 32 48 MLS.

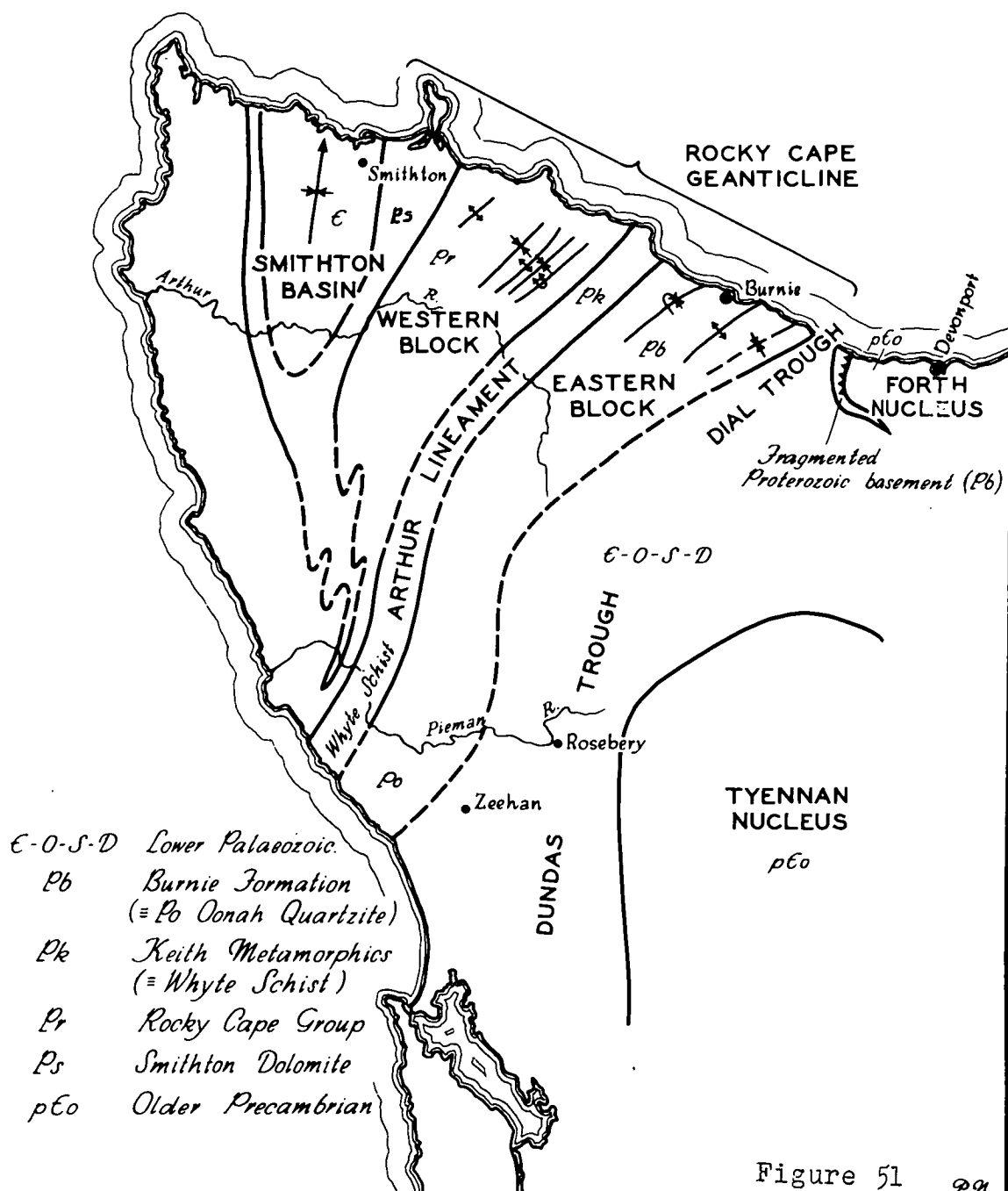


Figure 51 99.

Coming back to the structural framework in the west, the Proterozoic rocks of the Rocky Cape Geanticline continue all the way to the present west coast of Tasmania. This area is largely covered by the Smithton Dolomite, deposited in a broad shallow basin, and transgressing the Rocky Cape Group. This structural element is termed the Smithton Basin. Superimposed on this basin, along a meridional line through Smithton, is another belt of Cambrian sedimentary and volcanic rocks.

These major structural units are distinguished by their characteristic sedimentary assemblages and structural styles, and thus constitute the important palaeotectonic units of the Proterozoic-Lower Palaeozoic mobile belt in northwest Tasmania. Figure 52 is a composite profile showing the internal structure of the main elements, in the detail that is now known. It extends from the western block of the Rocky Cape Geanticline to the Forth Nucleus.

EVOLUTION OF THE ROCKY CAPE GEANTICLINE

The Rocky Cape Geanticline is virtually a discrete sedimentological and structural entity, and not merely a segment of the complex Precambrian basement that has been fragmented and uplifted during the lower Palaeozoic tectonic movements. The main features of the geanticline are the result of the Penguin Orogeny. The sedimentation phase is distinct in time

and effect from the deformation phase, rendering a relatively simple evolutionary picture.

Sedimentation phase

Spry (1962, p.125) and Solomon (1965) have previously advanced the picture of a sedimentary basin flanking the western side of the Tyennan Nucleus during late Precambrian times. This basin was probably part of a large geosyncline that may have extended as far as the mainland of present-day Australia. Rocks of similar lithology comprise the Proterozoic Adelaide System in South Australia.

A total of between 35,000 and 50,000 feet of sediment was deposited in this basin before the Middle Cambrian. The maximum at any one point may have been in excess of 30,000 feet. This amount of sedimentary rock rightly constitutes a geosynclinal pile, but it cannot be shown that the basin is a large elongate trough of geosynclinal dimensions.

Two distinct lithological assemblages are recognised in the Rocky Cape Geanticline, each representing different basins of deposition. The earlier basin formed with its axis approximately 80 miles to the west of the Tyennan Nucleus, and in this basin was deposited a suite of sediments that may be termed unstable-shelf. A thin veneer of stable-shelf sediments may have connected these sediments to the positive Tyennan Geanticline to the east

At least 20,000 feet of black shale, siltstone, orthoquartzite and minor dolomite, comprising the Rocky Cape Group, accumulated in the earlier basin. The base of this sequence is not exposed, and the earliest record of black pyritic shale indicates a quiet euxinic environment. There is only a minor indication of deposition by turbidity currents. It is possible that the finely laminated shales are varves, although there is nothing to suggest a glacial environment.

Deposition of enormous thicknesses of orthoquartzite sands followed the black shale, indicating a change in environment, although the extreme tectonic stability was maintained. A total of 9,000 feet of mature, well-sorted, quartz sand was deposited and reworked under shallow-water, free-circulating conditions, with some minor returns to the starved euxinic conditions that prevailed earlier. These orthoquartzites are remarkable for their complete lack of associated conglomerate. The orthoquartzites were derived ultimately from a distant granitic or gneissic source with no contribution from the Tyennan Nucleus to the east. It is probable that recycling of grains played an important part in the formation of the mature sands.

The paradox of great thicknesses of orthoquartzite must demand a delicate balance between subsidence and sedimentation over a long period of

geological time. Analysis of cross-bedding directions reveals unusual but simple diametrically opposed bimodal patterns, oriented transverse to the presumed basin elongation. These have been interpreted as transient expressions of a weakly oscillating palaeoslope on an otherwise flat, tectonically stable sea floor. In this way, the sands were continually reworked by zig-zaging along the axis of the trough under the influence of the transverse currents.

The first sign of emergence of the Rocky Cape Geanticline came at the conclusion of Rocky Cape Group sedimentation, when the axis of subsidence shifted closer to the Tyennan Nucleus. The rising geanticline was probably expressed only as a chain of islands just protruding above the sea. The embryonic geanticline was flanked immediately to the east by a deep, mildly active trough, and to the west by a more stable basin. At this early stage the effects of the Arthur Lineament first appeared. This line acted as a hinge about which subsidence occurred for Burnie sedimentation. A small basement high probably existed between the two major basins.

Following the tectonic quiescence during Rocky Cape Group sedimentation, the tempo of sedimentation quickened, and a thick sequence of flysch-type sediments accumulated in the new basin.

The dominant sedimentary regime was a slow accumulation of black shale, frequently and repetitiously interrupted by turbidity currents. Once again, most of the sediment appears to have been derived from an unknown and probably distant source. Some material came from the embryonic geanticline to the west, although this was a minor contribution. At isolated points there were minor outpourings of pillow lavas, probably of a spilitic nature. The association of pillow lava and graywacke is suggestive of mild eugeosynclinal conditions, however, the pillow lava is not sufficiently abundant to constitute an ophiolitic suite. The tectonic environment was still basically stable, there being only minor gravitational slumping.

The turbidity currents appear to have flowed parallel to the northeast - southwest elongation of the basin. Secondary transverse dispersal currents were also operating, which sifted and scoured the newly deposited sediment. There may also have been a small contribution from the older basement rocks to the east. Thus, at Penguin there is a cleaner paraquartzite which contains a proportion of quartzite grains. Also, near Goat Island, west of Ulverstone, there are conglomerate beds in the small segment of Proterozoic basement, in close proximity to the Forth Nucleus (Burns, 1964,

p.27). The stratigraphic positions of these more mature rocks are uncertain. They may represent a thin veneer of stable-shelf sediments linking the older Rocky Cape Group with the basement. Alternatively, they may represent a shallow-water facies of the normal flysch-type Burnie Formation. The structural evidence is not sufficient to distinguish between these two possibilities.

At approximately the same time as Burnie sedimentation, gentle subsidence was occurring on the other side of the embryonic geanticline. A thick dolomite sheet accumulated in this, the Smithton Basin. Although the stratigraphy is uncertainⁱⁿ the far northwest of Tasmania, the dolomite overlies several different rock types in the Proterozoic succession, and appears to be a regionally transgressive sheet. In the Black River area, there is a thin basal conglomerate which is interpreted as a shore-line deposit, formed during a slight transgression onto the rising embryonic geanticline. The conglomerate rests with angular discordance upon the Cowrie Siltstone, which is low down in the Rocky Cape Group. The conglomerate is composed of rounded orthoquartzite boulders identical to those occurring higher up on the Rocky Cape Group. It is clear, therefore, that there was a considerable amount of erosion off the embryonic geanticline. The angular

discordance of 20° recorded at Black River is probably due more to the initial dip of the conglomerate sheet rather than gentle tilting of the Rocky Cape Group.

The extent of the Smithton Basin is uncertain since the major portion of dolomite lies outside the area of study, and also much of the dolomite has been removed by Cainozoic erosion. However, it appears to have formed a vast sheet that blanketed much of the Proterozoic rocks in the far northwest of Tasmania. The Smithton Basin in Figure 51 is drawn from the present-day limits of the dolomite.

Deformation Phase

The structural phase of the evolution of the Rocky Cape Geanticline is the result of a regionally simple movement picture during which the two main sedimentary piles underwent folding and transport to the southeast toward the older Precambrian nucleus. The Rocky Cape Group was folded into a series of shallowly plunging folds trending northeast-southwest. At the same time the Burnie Formation was squashed against the older nucleus, forming large asymmetrical folds with belts of regional overturning. Detailed structural analysis in the eastern block reveals a complex picture of progressive deformation about approximately a constant axis. The Arthur Lineament, previously a hinge axis for sedimentation, became a zone of dispersed high-angle thrusting and

syntectonic metamorphism when the more competent western facies moved against, and partly overrode, the eastern facies. These movements constitute the Penguin Orogeny of Spry (1962, p.124).

There is a marked contrast in tectonic style between the structures of the western and eastern blocks of the Rocky Cape Geanticline. The western block consists of regular folds which are open and upright, becoming tighter and more asymmetrical toward the metamorphic shear zone. It appears, therefore, that the structure further to the west in the Smithton Basin is simple. The folds in the orthoquartzite horizons are usually broken by high-angle break thrusts. These faults appear to be deep-seated fractures that tapped a deep magmatic chamber allowing the intrusion of dolerite.

The length of arc of the folded Detention Sub-group in Figure 52 shows that it has been shortened from 22 miles to 17 miles. Most of this shortening occurs in the region between the Newhaven syncline and the metamorphic belt. Lateral shortening in stable-shelf type sequences is generally achieved by ideal concentric folding and low-angle imbricate thrusts which splay out from a plane of detachment at depth. This type of deformation may have occurred in the strongly crumpled and overturned zone close to the metamorphic belt.

Between Rocky Cape and the Newhaven syncline,

the projected form of the small folds in the Detention Sub-group suggests that the slight curvature of bedding is caused by movement along the faults rather than vice-versa. There is no obvious relationship between tight folds and axial-plane break thrusts, nor is there any relationship between the distance between thrusts and the thickness of the quartzite layer. These features, plus the high angle of the faults and the presence of dolerite within the fault planes, suggest the faults are deep structures which penetrated down through the basement at depth. It thus appears that the deformation in the western block may be controlled by segmentation of the older basement at depth. The movement in the basement was essentially vertical on high-angle thrust faults which penetrated up into the overlying sedimentary cover. This is similar to the concept of Belousov (1962), who termed this process, block folding.

On the other hand, the Burnie Formation deformed by simple buckling followed, or accompanied by, flattening. If the major structure of the Burnie Formation in Figure 52 is extrapolated to depth and then straightened out, it appears that it has been shortened by about 30%. This figure must be regarded as only approximate because of other factors such as the unknown thickness incorporated into the Keith Metamorphics and

the effect of flattening by pure shear.

The Burnie Formation is folded into the recumbent Coeee syncline, linked to the asymmetrical Round Hill anticline by a crumpled flat zone, and followed to the east by a syncline which brings the sedimentary pile back up, onto the Forth Nucleus. The end structure is the result of a progressive strain involving a sequence of five phases under approximately the same stress field. The first phase initiated the form of the major structure by a process of buckling with flattening in the steeply dipping limbs. Within these steep limbs, minor folds and slaty cleavage are well developed. Albite dolerite, similar to the syntectonic dolerite to the west, was intruded at this stage. These bodies are essentially sill-like, and were mainly intruded into areas devoid of minor folds of the first generation.

The second phase was of minor significance. During the third phase, further lateral shortening was achieved by buckling of the flat zone and a tightening of the early major folds. The fourth and fifth phases do not greatly contribute to the major structural configuration of the Burnie Formation.

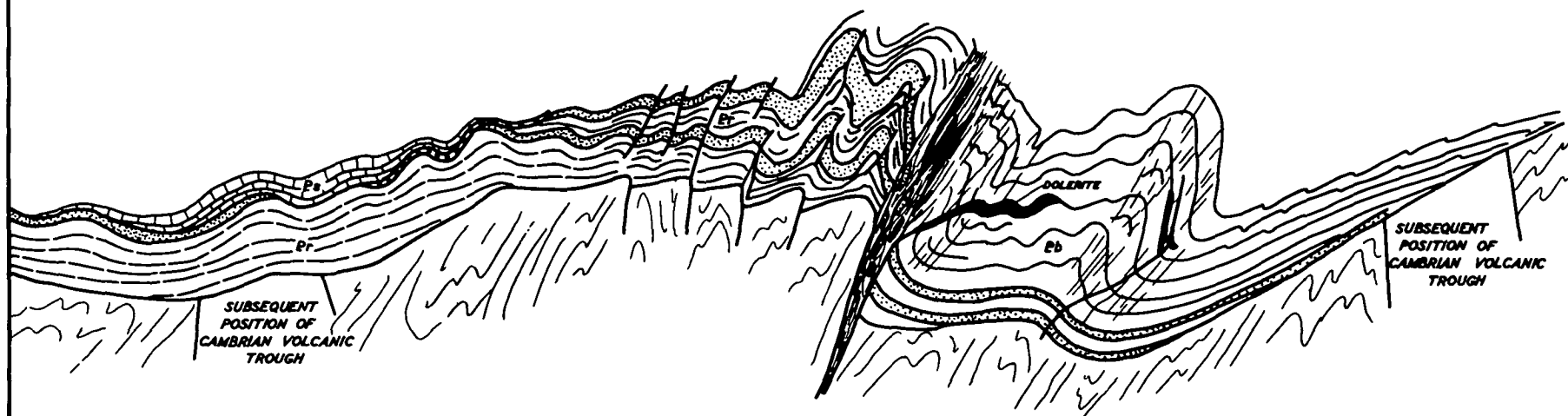
The axes of the main phases are oriented northeast-southwest. Because the phases are essentially concentric deformations, the intermediate principal stress

of the regional stress field must also have been oriented with a gentle plunge to the southwest. The principal maximum stress was therefore directed sub-horizontally in a northwest-southeast direction. The fact that some first generation minor folds have moderate or steep plunging axes does not demand a steeply plunging intermediate stress axis. This is due to a small rotation of about 30° of the intermediate axis in the horizontal plane after the first phase of folding. Continued folding of bedding to a vertical position about the new intermediate axis steepens the earlier fold axes. For example, when the 20° tilt on the b-kinematic axis for the Round Hill anticline is superimposed on top of the 30° obliquity between the P_1 and P_3 axes, a maximum plunge of 50° for the earlier fold structures in the Blythe Overturned Belt is formed.

Figure 53 is a palaeoprofile of the Rocky Cape Geanticline at the conclusion of the Penguin Orogeny. The large-scale lateral transport of the Burnie Formation to the southeast against the older basement supports the concept of Burns (1964, p.151) that the boundary between the upper and lower divisions of the Precambrian in the Goat Island area is an important low-angle thrust. This he called the Singleton Thrust. However, the age of the movement on the Singleton Thrust is considerably different to that envisaged by Burns,

W/VW

ESE



Er ROCKY CAPE GROUP
Pb BURNIE FORMATION
Pa SMITHTON DOLOMITE

**PALAEOPROFILE OF ROCKY CAPE GEANTICLINE
AT END OF PENGUIN OROGENY**

0 4 8 MLS.
APPROX.

who suggested it was related to his P2 deformation, (equivalent to P5 in this thesis). Any dislocation along this boundary must be due to the first or third generation movements in the Burnie Formation, or a combination of both.

It is during the structural development of the Rocky Cape Geanticline that the Arthur Lineament achieved its full importance. This line of weakness, which had been in evidence during the sedimentation phase, became a zone of intense shearing when the Rocky Cape Group moved against, and tended to override the Burnie Formation. The shearing was in a near-vertical direction with a west-side-up movement, producing a foliation parallel or sub-parallel to bedding. The dominant rock type within the metamorphic belt is albite schist with an assemblage indicating a lower-greenschist facies. The basic schist and amphibolite are considered to have been derived from sodic dolerite which was intruded into the shear zone early in the structural evolution. The structural analogy with the dolerite bodies to the west strongly suggests that they are metamorphosed equivalents of the Cooee Dolerite. This is supported by chemical evidence.

PALAEOZOIC HISTORY OF THE ROCKY CAPE GEANTICLINE

Following the major folding of the Penguin Orogeny, the crumpled sedimentary pile became a strongly emergent feature having essentially the same shape and internal structure as it now appears today. The Middle Cambrian segmentation of the Proterozoic basement that produced the Dial and Dundas Troughs may be related to this major emergence. The Cambrian trough was localised along the junction between the deformed Proterozoic belt and the older basement to the east, and thus can be thought of as a continuation of the proceeding tectonic pattern of continued shifting of the depositional axis closer to the Tyennan Nucleus.

Acid volcanics first accumulated in the Dial Trough (Burns 1964, p.153), and were followed by a eugeosynclinal suite of conglomerate, siltstone, keratophyric and spilitic intrusives, chert and graywacke, continuing up into Upper Cambrian times. Cambrian sedimentation of this association seems everywhere to be confined to the narrow trough, and did not lap very far onto the Rocky Cape Geanticline. The tectonic movements which terminated the Cambrian sedimentation (Jukesian Orogeny) had no noticeable effect on the internal structure of the Rocky Cape Geanticline. In western Tasmania, Solomon (1965, p.469) considered the dominant feature of the Jukesian Orogeny

was major faulting which uplifted the Tyennan and Rocky Cape Geanticlines. This type of tectonics does not appear to be so well expressed in northwest Tasmania (Burns 1964, p.161).

The picture may be somewhat similar on the western side of the Rocky Cape Geanticline where spilitic volcanics and upper Middle Cambrian sediments (Banks 1962, p.134, Solomon 1965) form a meridional belt through Smithton. This structure has been termed the Montagu Synclinatorium and thought of as a Tabberabberan structure (Solomon 1962, p.325), although it is more probably an expression of Jukesian faulting.

After the Jukesian Orogeny, Ordovician conglomerate, sandstone and limestone accumulated in a small shallow basin superimposed on the now quiet Dial Trough. The basal conglomerates transgress only slightly onto the Rocky Cape Geanticline at Penguin and Sulphur Creek. There is no information regarding the amount of transgression of the Ordovician limestone and later Siluro-Devonian miogeosynclinal sediments, but they may have formed a thin converging on the older structural units in the northwest of Tasmania. During, and following the Devonian Tabberabberan Orogeny, the Rocky Cape Geanticline was again emergent and the sedimentary cover was completely stripped off. However it is clear

that the geanticline as a whole did not participate in the Tabberabberan fold movements, being affected only in the fringe area of the Palaeozoic fold belt.

STRATIGRAPHIC AND STRUCTURAL CORRELATION WITHIN THE ROCKY CAPE GEANTICLINE.

The correlation of the Keith Metamorphics with the Whyte schist to form the Arthur Lineament, provides the basis for regional structural and stratigraphic correlations with the west coast fold belt. The Arthur Lineament stretches 80 miles between Granville Harbour on the west coast and Wynyard on the northwest coast, and is truly the "back-bone" of the Rocky Cape Geanticline. It is flanked to the northwest and southeast by two thick sedimentary assemblages, each corresponding to a different basin of deposition. In the lower Pieman River area, the Whyte Schist is flanked to the east by the Oonah Quartzite of large but unknown thickness. Because the structural trend of the lithological assemblages is parallel to the Arthur Lineament, it appears that the Oonah and the Burnie were deposited in the same sedimentary basin. The Oonah is lithologically similar but in places, is texturally more mature than the Burnie, and contains interbedded dolomite in the Zeehan and Dundas districts, (Blissett 1962, p.23). Spry (1964, p.45) has previously suggested this correlation on the grounds of lithological similarity although the supporting stratigraphic evidence that both the Oonah and the Burnie were at the

bottom of the Proterozoic succession is not correct. If this correlation is correct, it means that the Oonah is the youngest of the Proterozoic succession in the Zeehan-Corinna area. This broadly is the picture envisaged by Blissett (1962), who considered that the Oonah passed upward through the Crimson Creek Formation into the Dundas Group. The Success Creek phase of Solomon (1965, p.466), a transgressive phase of sedimentation which is post-Oonah and before the Cambrian volcanic activity, does not have a counterpart at Penguin.

To the west of the Arthur Lineament in the lower Pieman River area, the succession of Spry (1964) has similarities with the Rocky Cape Group. Both sequences contain thick beds of laminated shale, siltstone and mature quartzite, and once again, seem to have been deposited in the same basin. There are differences which prevent a direct correlation, formation for formation. For example, the sequence of Spry contains the Bernafai Volcanics (a formation of 1,300 feet of lava and tuff), and the orthoquartzitic Donaldson Group with conglomerate horizons. Spry (1964, p.45) has suggested the correlation of the Cowrie Siltstone with the Interview Slate, and the Rocky Cape Group orthoquartzites with the Donaldson Group. This latter correlation may be correct, but the possibility cannot be rejected that the Donaldson Group is equivalent to

the Forest Conglomerate, and thus represents the substantial amount of material that must have been eroded after the initial emergence of the embryonic geanticline.

To support the general configuration suggested here, Spry (1964, p.44) noted that the Oonah dips east, and off, the Whyte Schist, and the western sequence (Spry 1964, Figure 4) dips eastward toward, and under, the Whyte Schist. It is suggested that Spry's interpretation of the relationship of the western sequence to the Whyte Schist and the Oonah Quartzite is in need of revision.

Some correlations with McNeill's (1961) reconnaissance mapping of the middle Arthur River area can be made. The Neasey Quartzite and Slate, which flanks the Keith Metamorphics to the west, is the direct continuation of the Detention, Irby and Jacob Formations, taken together. In the absence of facings, McNeill thought that the Neasey overlies the metamorphic Keith Beds, and the Lawson River Siltstone overlies the Neasey. This succession must now be reversed so that the Neasey stratigraphically underlies the Keith Metamorphics. The Lawson River Siltstone would then correlate on stratigraphic on lithological grounds with the Cowrie Siltstone.

The dolomite blanket in the Smithton Basin

may well be the equivalent of the Savage Dolomite as suggested by Spry (1964, p.36), since they both appear to be transgressive over the Proterozoic succession west of the Arthur Lineament. It is unwise, however, to correlate all the dolomite horizons in the Proterozoic succession as suggested by Solomon (1965 p.466-467), especially in view of the dolomite horizon in the orthoquartzite suite of the Rocky Cape Group. This approach led Solomon to correlate the Smithton Dolomite with the Success Creek phase. The dolomite in the Success phase may well be of comparable time range to the Smithton Dolomite, but they were certainly deposited in different basins on either side of the already emergent Rocky Cape Geanticline.

PENGUIN OROGENY

All the tectonic movements which contribute to the internal structure of the Rocky Cape Geanticline are attributed to the Penguin Orogeny. Spry (1962, p. 124-126) defined the Penguin Movement as the tectonic event late in the Precambrian to account for the discordance and difference in structural history between the Burnie Formation (then included in the Rocky Cape Group) and the Dundas Group at Penguin. The dating of one specimen of Coee Dolerite at 700 m.y. (Spry, 1962, p.124) indicates a late Proterozoic age. This movement was shown to be pre-Middle Cambrian by Burns (1964).

The lower age limit is difficult to specify accurately. At Black River the folding is later than the deposition of the Smithton Dolomite. Spry (1962) has outlined the strong evidence that the Frenchman Orogeny is a period of widespread regional metamorphism separating older and the younger Precambrian rocks. Throughout this thesis, this concept has been followed, as it is believed that the sediments discussed here are of a younger Precambrian age. One strong line of evidence revealed by this study is that the Rocky Cape Geanticline is a sedimentological and structural entity within itself, more related to the Lower Palaeozoic tectonics of Tasmania than to the Precambrian basement.

In the Goat Island area, Burns (1964) suggested two phases for the Frenchman Orogeny in the Forth Nucleus, and found two phases to the Penguin Orogeny in the unmetamorphosed rocks. On extremely tenuous evidence, Burns (p.126) suggested that the second phase of the Frenchman may be equivalent to the first phase of the Penguin, and also that the second phase of the Penguin (P5 of this thesis) is equivalent to the dislocation movements on the Singleton Thrust. The present work does not support either of these correlations. It has been pointed out that P1 of the Penguin must be mainly responsible for the

dislocation movements on the Singleton Thrust, and Burns (p.152) shows that this dislocation post-dates the second phase of the Frenchman. Therefore both the two penetrative phases attributed by Burns to the Frenchman must be older than any deformation in the Burnie Formation.

The presence of the belt of metamorphic rocks within the Proterozoic succession has an important bearing on general concepts of Tasmanian Precambrian stratigraphy. Hitherto, the basic premise has been that the metamorphic rocks are older and separated from the unmetamorphosed rocks by the Frenchman Orogeny. This approach is not discarded and the arguments put forward by Spry (1962, p.121) are still considered to be generally correct, however it is clear that major stratigraphic subdivision in the Precambrian must be made on structural as well as metamorphic criteria.

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